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Foreland basins: lithospheric flexure, plate strength and regional stratigraphy

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FORELAND BASINS: LITHOSPHERIC FLEXURE, PLATE STRENGTH AND REGIONAL STRATIGRAPHY

A Dissertation

Submitted to the Graduate Faculty of the Louisiana State University and Agricultural and Mechanical College in partial fulfillment of the requirements for the degree of Doctor of Philosophy

In

The Department of Geology and Geophysics

by

John Londono
B.S., Univerisdad Nacional de Colombia, 1995
M.S., University of Alabama, 2001
August, 2004
To my family: Gilberto, Georgy, Patty, Lili, Anasofis and Raul Jr.........................

“Careless, Ironic, Violent
Is the way science want us, it is a woman, she will love just warriors”

F. Nietzsche
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ABSTRACT

Foreland basin subsidence through time is reproduced in this study, as the flexure of an elastic beam in an inviscid fluid under the vertical stress, caused by discrete-distributed loads. Thus, seismostratigraphic data from the Timor Sea peripheral foreland basin, in northwestern Australia, and the Putumayo retroarc foreland basin in the Colombian Andes, are forward modeled, at chronostratigraphic intervals, to assess the evolving geodynamic conditions of the basins. Results show that the accommodation in foreland basins varies as the depositional basement is vertically adjusted according to the regionally isostatic compensation of the lithosphere. Distributed tectonic (thrust belts) and sedimentary loads that act independently but consecutively during tectono-stratigraphic events, throughout the evolution of foreland basins, control the deflection of the plate that forms the foredeep of these depocenters. Accordingly, the loads limit the amount and distribution of available space for sedimentation. Results also reduce the role of eustasy to only 2 to 6% of the total accommodation, even in marine foreland depocenters. The strength of the plate remains invariable during the evolution of the basin at time scales of $10^6$ to $10^7$ m.y. Asymmetrical flexure, produced by oblique plate convergence, induces diachronous and local marine cycles at basin scale (100’s of km). Stratigraphic development of non-marine foreland basins is more likely to respond to the evolution of the equilibrium-profile during basin history.
CHAPTER ONE. INTRODUCTION

This dissertation explores the mechanical behavior of the lithosphere during the evolution of foreland basins and the effects of its deformation on regional stratigraphy. Through forward modeling of seismostratigraphic data from a peripheral foreland basin (Timor Sea) and a retroarc foreland basin (Putumayo, Northern Andes), the evolution of effective elastic thickness throughout time is estimated as well as the significance of supracrustal loads in controlling accommodation available for sedimentation.

Results in this study suggest that effective elastic thickness does not change at time scales of $10^6$ to $10^7$ m.y. in foreland basins. Tectonic (thrust belts) and sedimentary loads appear to control the amount and distribution of available space for sedimentation and facies, whereas eustasy seems to play a secondary role, independently of the tectonic setting of the basin. Therefore, accommodation in foreland basin varies as the base of the basin moves vertically, whereas the top (base-level), passively, respond to this changes. Asymmetrical flexure, produced by oblique plate convergence, induces diachronous and local marine cycles at basin scale. Stratigraphic evolution of non-marine foreland basins is more likely to respond to the process of adjustment of the fluvial system to the equilibrium profile (disrupted by tectonic loading and flexure) than to the previously suggested eustatic fluctuations (Weimer, R., 1986; Posamentier and Vail, 1988; Shanley, and McCabe, 1994; Emery and Myers, 1997).

Statement of the Problem

Foreland basins are sedimentary depositional spaces created primarily by the regional isostatic compensation of the elastic lithosphere to discrete loads (Beaumont, 1981; Jordan, 1981). The tectonic and mechanical evolution of foreland basins has been
well established theoretically (Beamount, 1981; Jordan, 1981; Pang, 1994). Today, it is recognized that these basins are formed as part of a mechanical entity composed of three basic regional elements: a bending plate, sedimentary deposits, and an orogenic belt. The vertical movement of the basement can generate or limit the accommodation for sedimentation; consequently, these movements can be stratigraphically recorded.

Extensive work has been carried out on diverse aspects of foreland basin development (e.g. Beamount, 1981; Stockmal et al., 1986; Cloetingh, 1988; Jordan and Flemings, 1989; Coakley, 1991, Peper, 1993; Pang, 1994; Shanley and McCabe, 1994; Cardozo and Jordan, 2001). Yet, some aspects regarding the puzzle that comprises the vertical plate movement remain unexplored. Via forward modeling of the seismostratigraphic record, this work will attempt to answer the following questions:

(1) Does the strength of the continental lithosphere change through time during the evolution of a foreland basin. What causes this change? What is the time scale of these changes? Does any evidence of these changes exist in the stratigraphic records?

(2) What is the amount of subsidence caused solely by tectonic loading in foreland basins? Is this subsidence of the same order of magnitude of that caused by sediment and water loading in foreland basins? Is it really possible to quantify them from the stratigraphic record?

(3) Are geodynamic conditions, such as plate strength, plate deflection, inelastic yielding and regional (basin and stratigraphic) geometry, similar in peripheral and retro-arc foreland basins? If not, does the regional tectonic setting have any impact on such conditions?
Modern forward numerical models for stratigraphic data reduction have provided an analytical tool for understanding the tectonic evolution and the basin fill progression of numerous foreland basins (Lerche, 1990). The integration of quantitative data of sediment supply and denudation, sea level fluctuations, lithospheric flexure, load geometry and sub-surface processes (Flemings and Jordan, 1989; Mitrovica et al., 1989; Jordan and Flemings, 1991; Reynolds et al. 1991; Peper, 1993), and the examination of variable rheology for the lithosphere (Beaumont, 1981; Stockmal et al., 1986) lead to the development of deterministic models of basin evolution for prediction of facial architecture, thermal gradients, amount of regional and local isostatic compensation and gravity anomaly, among others factors.

**Dissertation Outline**

Basin subsidence analysis begins with data reduction to initial depositional conditions (decompacted thickness, Appendix 3). Modeling of flexure produced by sediment loading, tectonic loading and water loading (theory in Appendix 1) is then carried out at chronostratigraphically significant intervals. Therefore, it is possible to assess temporal changes in rheological conditions (Appendix 1). For flexural modeling, a computer routine developed in MatLab® incorporates distributed loads for sediments, tectonic wedges using the Coulomb criteria and eustatic fluctuations (Appendix 4). Even though the Timor and Putumayo foreland basins are developed at convergent margins, the thermal age of the lithosphere in both cases exceeds 300 m.y., therefore subsidence related to thermal cooling is negligible in both basins. The effects of intraplate stress fluctuation are not considered. They are short-term fluctuations (Cloetingh et al. 1989) at least two order of magnitude less than the resolution of our data. Additionally they
become significant in weak lithospheres i.e. ~40 km of effective elastic thickness or less (Cloetingh et al., 1989; Peper, 1993).

Chapter 2 is a paper presented at the American Geophysical Union fall meeting in 2001. The paper is in press for the Tectonophysics Special Issue on “Ophiolites and continental margins of the Pacific Rim and the Caribbean region”. This paper evaluates the geodynamic evolution of the Timor Sea foreland basin. Two stratigraphic intervals divided by a regional unconformity are treated using the procedure described above. Only 2-D linear load flexural analysis is carried out. Results indicate that effective elastic thickness of the continental Australian lithosphere lies between 80 to 100 km and did not change during the evolution of the basin. The late Tertiary collision between Eurasian and Australian plates in Timor Sea was oblique, and proceeds from southwest to northeast. The down-warping of the plate produced normal faulting whose timing and spatial distribution results from the oblique nature of the collision.

Chapter 3 uses the database of chapter two. A more detailed seismic interpretation is carried out and marine transgressive-regressive cycles are identified. Flexural analysis, using the MatLab routine with distributed sediment and tectonic loads, is completed, and subsidence is adjusted for eustasy (assuming local isostatic compensation). Results indicate that the oblique collision between Eurasian and Australian plates produces an asymmetric deflection of the Australian continental lithosphere that causes differences in the number of marine cycles between the southwest (seven cycles), where collision began, and the northeast (five cycles) to where collision propagated later. Effective elastic thickness is between 70 and 100 km and does not change at sequence-time scale. Tectonic loading, produced mostly for the Timor Island accretionary prism, is responsible
for up to 70% of the total subsidence. Sediment load is responsible for about 28% of total subsidence and eustasy for the remaining 2%.

Chapter 4 studies the Putumayo retro-arc foreland basin along the Colombian Andes. Seismic and well data from oil industry are used to assess the flexural history of the basin. Distributed tectonic and sediment loads are modeled using the MatLab routine. The non-marine foreland sedimentary succession is divided into four tectonostratigraphic units by long-term regional hiatuses. Results indicate that the effective elastic thickness varies between 20 and 40 km and did not change at the scale of $10^7$ m.y. Sediment loading is responsible for up to 78% of the total subsidence and tectonic loading is responsible for the remaining 22%. The deflection of the plate is mostly symmetric and reaches subsidence rates up to 140 m/m.y. Seismic expression of this non-marine foreland basin, shows dominantly onlap reflector-geometry, with basinward shifts that indicate deepening of the basin. A conceptual model about development of continental sequences is presented. It includes flexural compensation and the concept of graded river. Hiatuses are interpreted as developed during periods of equilibrium in fluvial systems. Sea-level appears to have no influence in creation of accommodation and regional geometry of the foreland basin.

References


CHAPTER TWO. GEODYNAMICS OF CONTINENTAL PLATE COLLISION DURING LATE TERTIARY FORELAND BASIN EVOLUTION IN THE TIMOR SEA: CONSTRAINTS FROM FORELAND SEQUENCES, ELASTIC FLEXURE AND NORMAL FAULTING

Introduction

Foreland basin subsidence is primarily related to the vertical deflection of the lithosphere caused by loading of orogenic belts, although subsurface loads, and sediment and water bodies also play an important role in the process (Beamont, 1981; Watts, 2001). The distinctive wedge-like geometry of foreland basins is a direct result of asymmetric subsidence caused by the accumulation of vertical stresses toward the deeper end of the plate. The stratigraphic record complies with the regional architecture and therefore represents the geometry of the basin at depositional time (Beaumont, 1981; Turcotte and Schubert, 1982; Kruse and Royden, 1994; Dorobek, 1995; Yang and Dorobek, 1995). The timing, amount of vertical force and total plate deflection may be determined via flexural modeling, using the stratigraphic record deposited during tectonic loading (Kruse and Royden, 1994; Cardozo and Jordan, 2001). The basin subsidence can be modeled as the flexural response of an elastic plate to a vertical linear load (Turcotte and Schubert, 1982). Changes in subsidence across the strike of the basin through time document the amount (Cardozo and Jordan, 2001) and polarity of loading, and thus the variation in lithospheric strength. Normal faulting has been reported along the subducting plates as a result of the flexing lithosphere undergoing inelastic deformation (e.g., Bradley and Kidd, 1991). Extensional faulting is expected to affect areas of high curvature within the upper half of the bending plate. Variations in the amount of tectonic transport of the overridden plate through time change the amount of bending and the
spatial distribution of associated normal faults. The forebulge, a lithospheric protuberance developed as an elastic effect of the deflection on the distal part (landward) of the bending plate, is expected to reach highs of up to hundreds of meters and to extent over hundreds of km in width. It is recognized by unconformities and sedimentary pinch-outs on both of its flanks (e.g. Dorobek, 1995).

The late Tertiary foreland basin that developed in the Timor Sea (Fig. 2.1) during the latest episode of collision between the Australian and Eurasian plates (~ 6.5 to 1.6 Ma.) provides an excellent example of a peripheral foreland basin (Miall, 1995; Lorenzo et al., 1998; Tandon et al., 2000). The Timor Trough, a deep depression, is created by the deflection of the Australian plate under the load of the accretionary wedge of the Banda orogen (Hamilton, 1979; Harris, 1991), as well as sediment loading (Lorenzo et al., 1998). Tandon et al. (2000) used sea floor bathymetry and gravity data to model the present lithospheric strength of the Timor Sea foreland basin. The model represents the geometry of the Timor Trough as a ~300 km wide, ~2000 m deep depression with a 300 m high forebulge (Fig. 2.2). The Australian continental shelf has been heavily affected by reactivation and/or new growth of normal faulting during foreland evolution. Lorenzo et al. (1998) emphasize the role of normal faulting as evidence of inelastic yielding during bending of the Australian plate. O'Brien et al. (1999) recognize the role of plate downwarping in the reactivation of Jurassic faults and formation of Mio-Pliocene fault arrays. Some of these faults are still active, as shown by vertical displacement of the present sea floor (Lorenzo et al., 1998).
Figure 2.1. (A) Tectonic setting of the Timor Sea. A-A’ cross-section in figure 2. (B) Seismic and well data. Line number as referred by AGSO. Wells: (1) Ashmore Reef 1; (2) Delambre 1; (3) Buffon 1.
We use flexural models representing the bending-plate to estimate the effective elastic thickness (Appendix 1) during evolution of the basin. We calculate the amount of loading, infer the nature of the collision between the Eurasian and Australian plates and characterize how the Australian lithosphere has responded to the load of the Banda Orogen accretionary prism.

The position of the geodynamic elements of the bending-plate model during basin evolution provides a tool to reconstruct the regional basin history in terms of subsidence (deflection), tectonic transport (linear load position) and the extent of the Australian lithosphere flexed by loading. We also evaluate the impact of inelastic yielding and of the curvature caused by vertical loading on plate strength. The timing of the apparent displacement of new or reactivated faults developed during the evolution of the foreland basin is used as an indicator of the polarity of collision between the Eurasian and Australian plates. Our results, based mainly on extensive seismic data, indicate that despite continuous deformation through time, no significant change in lithospheric strength has occurred during foreland development. We also recognize NNE oblique nature of the plate collision in the Timor Sea.

The amount of deflection and inelastic deformation in the southwestern part of the Timor Sea reveals that this area has undergone deformation and has been heavily affected by tectonic loading since the Late Miocene time. In contrast, in the northeastern region the effect of loading has been more substantial since the Late Pliocene time.
Tectonic Setting

The northwestern Australian continental plate is presently in steady-state collision with the Banda arc and Timor Island (Hamilton, 1979; Fig. 2.1). The colliding process began to affect the Australian North West Shelf in the Timor Sea in late Miocene time. The N20E convergence of the Australian plate is estimated at a rate of 71 mm/y (Tregoning et al., 1994). However, GPS studies indicate that the Australia-Timor convergence has ceased and the Australia-Eurasia collision is accommodated by back-thrusting north of Timor along the Flores and Wetar thrusts in the Banda Arc (Genrich et al., 1995).

Figure 2.2. Present Timor Sea/Timor Trough flexural model, modified from Lorenzo et al., (1998). Extent of effective plate from node point (point of zero deflection and marks the beginning of the forebulge) is ~ 230 km for continuous model and ~155 km for broken plate. The flexural parameter (Alpha) is ~ 98 km. (see Figure 2.1 for location).
The Australian lithosphere has been subducted under a 200 km wide zone of Eurasian lithosphere (Richardson and Blundell, 1996), including Timor Island, an accretionary prism (Audley-Charles, 1986) developed over part of the stretched Australian continental crust. The Australian North West Shelf in the Timor Sea has experienced various tectonic episodes since Paleozoic times. O'Brien et al. (1996) note that NE-SW trending structures (Vulcan and Malita grabens, Fig. 2.1) reveal late Devonian-early Carboniferous rifting events, in contrast to the NW-SE trend of Jurassic rift-related basins (Pretelt sub-basin, Sahul Syncline, Fig. 2.1). Hamilton (1979), using refraction velocities, reports more than 4 km of sedimentary cover overlying continental crust in the area towards the toe of the tectonic wedge. Continental crust with decreasing thickness, from 35 km in the Kimberly Highlands to 26 km under the outer shelf near the Timor Trough (Petkovic et al., 2000), underlies the Australian shelf. Petkovic et al. (2000) estimate the attenuation of Precambrian basement rocks from 35 to 13-14 km across the margin ($\beta=2.6$).

**Foreland Stratigraphy**

Since the late Miocene, the Northwestern Australian Shelf in the Timor Sea has been a carbonate ramp whose stratigraphic architecture has been driven principally by sea level fluctuations of diverse origin, such as tectonic subsidence and eustasy (Apthorpe, 1988). Two unconformities have been identified in the foreland succession (Fig. 2.3). The oldest one, at the base, interpreted from the juxtaposition of shallow and deepwater facies, was identified in the commercial wells Delambre 1 and Buffon 1 (Fig. 2.1; Aptorphe, 1988).
This regional unconformity, at the top of Middle Miocene limestones (planktonic foraminifera of Zone N10), is overlain by late Miocene (Zone N15) deep-water carbonates. The second unconformity, reported by Veevers et al. (1974) at the ODP Leg Site 262, separates shallow water Upper Pliocene dolostones and calcarenites from folded shallow-water Lower Pliocene carbonates, at the axis of the trough.

Boehme (1996) also identifies this unconformity along proprietary seismic profiles. According to Hillis (1992), the hiatus represents a short-term depositional break (< 1 m.y.) at Ashmore Reef 1 well (Fig. 2.1). The top of the foreland deposits are punctuated by the youngest lithified sediments dated at ~1.6 Ma (Apthorpe, 1988). The lithology of the entire succession, poorly known, is described in some wells as greenish-gray calcareous lutites and siliceous siltstones of outer shelf, shelf edge and platform environments (Apthorpe, 1988).

The seismic reflectors that represent the top of late Pliocene and the unconformities at the base of late Miocene and late Pliocene are identified within our seismic data. The interpretation (Fig. 2.3) was carried out by integration of well information (sequence thickness and age) and previous reports (Ostby and Johnstone, 1994; Woods, 1994; Wormald, 1988) based on seismic data (reflector geometry, interval time and thickness). Since our goal is to analyze the temporal evolution of the foreland basin, we divide the foredeep succession in two packages, as does Boehme (1996). Using the regional unconformities, we name the oldest package Sequence A (late Miocene to early Pliocene) and the youngest package Sequence B (late Pliocene). The present combined thickness of these sequences varies between 20 and 800 m as estimated from well data and seismic time-depth conversion curves.
Figure 2.3. (A) Uninterpreted Seismic Line 118_15 (see Figure 1 for location). (B) Interpreted seismic profile. Note the characteristic regional wedge-like geometry of the foreland succession (highlighted by black lines). Interval-velocity range is between 1500 and 2200 m/s. The boundaries of the sequences are short-term unconformities. Normal faults were developed during plate bending (~6.5 to Recent) and older normal faults were reactivated.
Data

Over 2000 km of two-dimensional seismic reflection data, 4 to 6 seconds recording, 48-fold coverage, from 192 channels with 50 meters shot interval and 12.5 meters common-depth points (CDP's) from the Australian Geological Survey Organisation (AGSO) 1996 seismic program (Fig. 2.1) are used as the primary source of data. Most shiptrack lines are parallel to the SSE Miocene-to-Recent tectonic transport direction. Although seismic profiles only sample the Australian shelf and southern Timor Trough, they contain enough information for flexural modeling. Additionally, well log information and seismic data from the literature allow us to tie biostratigraphic and lithological data to the seismic record (Figs. 2.1 and 2.3).

Methods

Sequence boundary unconformities are mapped within the entire seismic data. Areas like the Cartier Trough, where the interpretation is ambiguous due to high deformation, are omitted from subsequent flexural analysis. The largest uncertainty in the reconstruction of decompacted thickness and flexure profiles is associated with estimates of sequence thickness calculated using interval velocities derived from the seismic semblance analysis. Using stacking velocities and following Telford et al. (1990), interval velocities were estimated between 1500 and 2500 m/s (TWTT). We estimate the time-depth conversion using these velocities for Sequence A and B. Time was taken from the seismic sections. According to the interval velocity range, the error in thickness calculation might be 11 to 14 %. The results, where control is possible, match the thickness range constrained by well information. Isopach maps derived from our seismic
interpretation (Fig. 2.4) show local depocenter distribution within the general deepening of the basin during foredeep sedimentation.

**Flexural Models**

Since there are not enough data to constrain the water depth along the profile at any time, we assume that the accommodation in the evolving Timor Sea foreland basin is completely filled with sediments throughout the two depositional periods we are analyzing. If an underfilled basin is assumed, the upward isostatic force (Appendix 1) has two components: one is produced by the sediment infill and the other is produced by the water infill. The deflection caused by the linear load will decrease with increasing amount of sediment. Using the parameters in Table 1, the difference of maximum deflection between a basin full of sediments and a basin full of water is between 14 and 16%. We model the deflection caused by the tectonic loading and the weight of sediments calculated from the record of each cycle but keeping in mind that today’s Timor Trough contains a water column of about 2500 m (Hamilton, 1979). Thus, we take the base of the sequence profile, at a particular period, to represent the down-warping top of the lithosphere and the top of the profile as a horizontal surface. This, ideally flat surface, rarely coincides with sea level; rather, it is an abstraction that represents a horizontal datum of zero deflection and marks the position of node point along the profile, as well as the beginning of the forebulge (Fig. 2.2). These profiles are used as the principal constraint to develop models of plate deflection. The theoretical curves, produced primarily by variation in effective elastic thickness and in amount of loading, must match adequately the geometry of the data (decompacted thickness profiles) and fit into the geological model of the basin. We reproduce the theoretical profiles following
the mechanical model of the two-dimensional flexure of an elastic beam, of constant 
elastic thickness, lying over a viscous substratum (Turcotte and Schubert, 1982). Table 
2.1 summarizes the parameters used in these calculations. Seven seismic profiles were 
chosen to carry out flexural modeling (Fig. 2.1). We discard lines crossing areas 
deformed by salt-tectonics and consider only seismic lines collected parallel to the 
regional NW-SE tectonic load transport direction.

Foreland basin studies use both broken and continuous plate models. For 
eexample, Tandon et al. (2000) use a broken plate for modeling the Australian Shelf, 
whereas Kruse and Royden (1994) use a continuous plate model for the Apennine and 
Dinaride foreland basins in the Adriatic Sea. Using the present bathymetric profile of the 
oceanic floor in the Timor Sea (Fig. 2.2), we test both continuous and broken plate 
models following Turcotte and Schubert (1982) and geodynamic values shown in Table 
2.1. Broken plate seems to be more appropriate given previous interpretations of the 
extent of the continental Australian lithosphere only tens of miles north of Timor Island 
(Hamilton, 1979; Lorenzo et al., 1998).

It is important to consider the position of our data within our modeled profiles at 
each particular time (Figs. 2.1 and 2.2). Our study area undergoes NNW displacement 
with respect to the Eurasian plate during collision (Hamilton, 1979). The same well 
location, for example, will over time occupy a point closer to the trench. Consequently, 
within each profile the seismic data partially records at least two independent flexed 
stages of the plate. Sequence A contains sediments deposited more distally from the 
trench axis, while Sequence B contains sediments deposited at sites more proximal to the 
trench axis. Seismic data are assumed to be located at the most proximal position to
Timor Island (tectonic load). Therefore, the present distances between the point of zero deflection (node point, Fig. 2.2) and the position of the linear load (Point of maximum deflection Xo, Table 1, Figs 2.5 and 2.6) must be the minimum acceptable value for our models, assuming that the plate remains homogeneous throughout time.

Table 2.1. Summary of the mechanical parameters used in flexural models. The governing equation for the bending plate is \( D\left(\frac{d^4w}{dx^4}\right) + (\rho_m - \rho_w) gw = 0 \), (Turcotte and Schubert, 1982) assuming a linear load, where \( D \) is flexural rigidity, \( w \) is the deflection, \( \rho_m \) is the density of the mantle, \( \rho_w \) the density of water infill, \( g \) gravity acceleration. Bending moment (M) is given by equation \( M= D\left(\frac{d^2w}{dx^2}\right) \). Curvature is given by equation \( R= M/D \) (Turcotte and Schubert, 1982).

<table>
<thead>
<tr>
<th>Definition</th>
<th>Symbol</th>
<th>Value/Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density of Water</td>
<td>( \rho_w )</td>
<td>1030 kg/m³</td>
</tr>
<tr>
<td>Density of Mantle</td>
<td>( \rho_m )</td>
<td>3300 kg/m³</td>
</tr>
<tr>
<td>Flexural Rigidity</td>
<td>( D )</td>
<td>Nm</td>
</tr>
<tr>
<td>Effective Elastic Thickness</td>
<td>( Te )</td>
<td>km</td>
</tr>
<tr>
<td>Flexural Parameter</td>
<td>( \alpha )</td>
<td>m</td>
</tr>
<tr>
<td>Maximum Deflection</td>
<td>( W_o )</td>
<td>m (At point Xo)</td>
</tr>
<tr>
<td>Gravitational Acceleration</td>
<td>( g )</td>
<td>9.8 m/s²</td>
</tr>
<tr>
<td>Young Modulus</td>
<td>( E )</td>
<td>( 11 \times 10^{11} ) Pa</td>
</tr>
<tr>
<td>V Poisson Ratio</td>
<td></td>
<td>.25</td>
</tr>
<tr>
<td>Load (Linear)</td>
<td>( P_o )</td>
<td>N/m</td>
</tr>
<tr>
<td>Bending Moment</td>
<td>( M )</td>
<td>N</td>
</tr>
<tr>
<td>Curvature</td>
<td>( R )</td>
<td>1/m</td>
</tr>
</tbody>
</table>
Figure 2.4. Isopach maps for late Miocene-early Pliocene (A) and early Pliocene-late Pliocene (B). Note the increasing thickness toward the trench. Map A shows some regional tectonic structures like the Sahul Syncline and Vulcan Sub-basin. In Sequence B (map B) structure are more subtle from the isopach map.
**Decompaction Calculation**

We assume that compaction is the product of decreasing porosity due only to the mechanical non-reversible process of expulsion of pore water. Sediment decompaction is carried out to obtain the thickness at the time of deposition. Appendix 3 explains in detail the method we follow for estimates of porosity and decompaction. Values for c, the porosity-depth curve-slope coefficient (0.57 km\(^{-1}\)); and initial porosity, (0.63), are taken from studies based on well-logs in the Australian platform (Hillis, 1990).

From the thickness error introduced by interval velocities inaccuracy, porosity can change up to ~3%. This in turn introduces an error in decompacted thickness up to ~7%, which we consider an acceptable value since the average thickness change is ~28%.

**Flexure-Related Normal Faulting**

Reactivated normal-fault planes dip between 33 and 40 degrees in the shallow section within post Cretaceous sediments. In the deeper section, some faults are listric and their traces die out in pre-Mesozoic strata, irrespective of fault vergence, as shown by Lorenzo, et al. (1998). A regional detachment zone is not easily distinguishable along the seismic data. Nonetheless, Paleozoic evaporite beds are the most probable detachment level in the area. O'Brien et al. (1993) report some of these faults cutting basement at 15 to 20 km depths. Newer faults, developed during foreland evolution, die within Cretaceous or younger strata and dip between 40 and 55 degrees along the entire fault trace.

We assume that the fault slip measured along the fault plane in the foreland basin deposits indicates a comparative amount of deformation caused by extension due to flexure. Normal faults caused by plate bending are expected to develop in areas of high
curvature. The more flexed the plate is, the higher the curvature, where greater tensional stresses produce brittle deformation (Bradley and Kidd, 1991).

To calculate the apparent displacement in the section along the fault plane in both sequences, vertical stratigraphic separation is measured in meters and corrected by the angle of the fault. The apparent slip separating the reflector representing the top of Sequence A documents not only the displacement of that particular period of deformation, but also subsequent movements. Since we are interested in the apparent slip developed during Sequence A time, it is necessary to subtract any slip up-dip of the fault plane affecting younger deposits. We assume that the fault slip is homogeneous along the entire segment affecting foreland strata. We identify and measure over 200 new or reactivated normal faults active during plate deflection.

**Results**

**Flexure Models**

Subsidence of the late Tertiary foreland basin in the Timor Sea is modeled by flexural deformation of a homogeneous, elastic, continuous plate. During Sequence A time (~6.6 -~3.4 Ma) the maximum decompacted thickness is calculated as ~1200 m along seismic profile 116_04, and the minimum is ~ 400 m along profile 165_09 (Fig. 2.5). However, these values are not a direct measure of the deflection of the plate. In order to do that, it is necessary to evaluate the position of our data and best fitting models within the regional deflecting trend. A comparison between best-fitting curves indicates that during Sequence A time, the western part of the Timor Sea was undergoing remarkably more deflection at the end of the effective plate than the eastern part under similar elastic thickness. In contrast, during Sequence B time (~3.4 - ~1.6 Ma) a
comparison of flexural models indicates that the east Timor Sea was undergoing more subsidence (flexure) than the western part. The deposits are thicker in the eastern part of seismic control (up to ~979 m) along lines 118_15 and 118_02 (Fig. 2.6). During this time, the decompacted thicknesses coincide with predicted values of plate deflection. A range of effective elastic thickness values (20 to 120 km) is tested, but only cases where values lie between 80 and 100 km appear to match the decompacted stratigraphic data. For the base of Sequence A in the western-central part of the survey (lines 165_09, 163_01_15, 116_04), effective elastic thickness values between 80 and 100 km produce acceptable fitting curves to the data. In the easternmost survey area, in turn, only an effective elastic thickness of 100 km appears to fit the seismic data in continuous plate models. Our analysis suggests a maximum deflection of the Australian lithosphere of about 3500 m in the west Timor Sea (about 2500 m at the axis of the Trench). Table 2.2 summarizes the results of the best fitting curves for both sequence times.

Flexural model results are limited due to the parameterization of input data. Linear load for example, is used here, however, it represents the combined effect of the vertical stresses induced by tectonic and sediment loads. The geometry of the load will change the geometry of the resultant deflection. Another factor limiting our models is the extent of the data. We have control along 100 to 240 km. However, the deflection involves 450 to 500 km of lithosphere. Still, the lateral continuity given by seismostratigraphic data is less speculative than well data for example. Additionally, the Timor Sea foreland basin is heavily affected by normal faulting related to salt tectonics and flexure. This deformation creates changes in sequence-thickness not related directly to the down-warping of the plate. These changes, at times, are difficult to reconcile with
flexural models. Some lines, therefore, were not used for modeling.

Table 2.2. Flexural model summary. Variables as defined in Table 2.1. $X_0$ is the distance between the linear load at the position of maximum deflection and the node point or point of zero deflection. $X_m$ is the position of the maximum bending moment along the model.

<table>
<thead>
<tr>
<th>Sequence A</th>
<th>EET (km)</th>
<th>$D$ (Nm)</th>
<th>$\alpha$ (km)</th>
<th>$X_0$ (km)</th>
<th>$P_0$ (N/m)</th>
<th>$W_0$ (m)</th>
<th>$X_m$ (km)</th>
<th>$M$ (Nm)</th>
<th>$R$ (m)</th>
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<td>4.6E24</td>
<td>168</td>
<td>397</td>
<td>2.7E13</td>
<td>3552</td>
<td>265</td>
<td>2.35E17</td>
<td>5.11E-8</td>
</tr>
<tr>
<td>163_01_15</td>
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<td>9.0E24</td>
<td>199</td>
<td>469</td>
<td>2.2E13</td>
<td>2400</td>
<td>313</td>
<td>2.28E17</td>
<td>2.52E-8</td>
</tr>
<tr>
<td>116_04</td>
<td>80</td>
<td>4.6E24</td>
<td>168</td>
<td>397</td>
<td>1.3E13</td>
<td>1940</td>
<td>265</td>
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<td>469</td>
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</tr>
<tr>
<td>118_15</td>
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<td>199</td>
<td>469</td>
<td>8.5E12</td>
<td>928</td>
<td>313</td>
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<th>$D$ (Nm)</th>
<th>$\alpha$ (km)</th>
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<th>$P_0$ (N/m)</th>
<th>$W_0$ (m)</th>
<th>$X_m$ (km)</th>
<th>$M$ (Nm)</th>
<th>$R$ (m)</th>
</tr>
</thead>
<tbody>
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<td>1.0E13</td>
<td>655</td>
<td>313</td>
<td>6.21E16</td>
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<td>2180</td>
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<td>3000</td>
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<td>2.9E17</td>
<td>3.21E-8</td>
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</tbody>
</table>
Figure 2.5. Flexural model for Sequence A time (~ 6.5 - 3.4 Ma). Gray line represents the seismic data (decompacted thickness). Black line represents the theoretical models. These figures represent the best match between both curves. EET is effective elastic thickness. The linear load calculated for each model is located at the most deflected end of the profiles, the end of the effective elastic plate (See Appendix 1 for a complete explanation). Note the extent of the deflection between 400 and 470 km.
Figure 2.6. Flexural models for Sequence B (~3.4-Recent). Note the homogeneous effective elastic thickness.
Normal Faulting

The tension in the upper-half plate caused by deflection, produced abundant normal faulting in the Australian Platform (Figs. 2.3 and 2.7). A greater number of faults with higher displacement are found in late Miocene to early Pliocene than in late Pliocene times (Fig. 2.8a). The distribution of these faults suggests that the stresses caused by vertical loading were probably higher in the west than in the east Timor Sea, as also shown by estimates of curvature (Table 2.2, Fig. 2.7). The apparent fault slip measured along the deflected plate shows that the mean displacement during the Sequence A period was 40 meters in the western area (lines 165_06, 165_09 and 161_03_15) and only ~24 meters in the eastern area (lines 118_02, 118_15 and 118_06). Therefore, the average slip is ~66% higher in the west than in the east. The maximum curvature for this period, which occurred toward the west, is ~5.1x10^{-8} m. During Sequence B time, normal faulting displacement was also higher in the west Timor Sea. The mean slip is 34 meters in this area (lines 165_06, 165_09, and 161_03_15), whereas for the east Timor Sea it is < 25 meters (lines 118_02, 118_15 and 118_06).

The average displacement of these faults is up to 36% higher in the west than in the east (Fig. 2.8). However, the maximum curvature during this period (~3.4 – 1.6 Ma) occurs in the east Timor Sea (Table 2.2) and it is estimated as ~3.2x10^{-8} m. Individual fault-slip measurements (Figs. 2.7 and 2.8) show that the absolute amount of displacement is up to two times higher in the west Timor Sea (120 meters) than in the east Timor Sea (maximum slip of 50 meters).
Figure 2.7. Normal flexural-related faulting for a section of (A) western line 163_01_15 and (B) eastern line 116_04. Note the early-late Pliocene sequence almost unaffected in the western line (A).
Figure 2.8. (a) Fault slip frequency and temporal distribution. Note the higher displacement during Sequence A time than during Sequence B time. In (b) and (c) note the distribution of fault slip along the east and west Timor Sea during foreland evolution.
Discussion

Variations in the effective elastic thickness of the Australian lithosphere in the Timor Sea are not evident during development of the foreland basin. McNutt (1984) shows that there is a correlation between thermal age and plate thickness: plates older than 100 m.y. are expected to be strong (i.e. high rigidity). By contrast, Watts (2001) notes that no simple relation exists between thermal age and continental lithosphere rigidity. Rather, he favors the role of the crust composition in the present-day geothermal gradient as the key factor controlling the continental effective elastic thickness. Since the last tectonic event to affect the Australian plate prior to bending was Triassic-Jurassic rifting (geothermal age), the Australian continental lithosphere along the northwestern shelf is expected to be strong.

Our results show that at the beginning of continental plate collision, during Sequence A time, models favor an effective elastic thickness ranging from 80 to 100 km. The Australian lithosphere in the east Timor Sea appears consistently strong (100 km of effective elastic thickness), while in the central area of the survey (Line 116_04, Figs. 2.1 and 2.5) and in the westernmost region of the seismic survey (line 165_09, Figs. 2.1 and 2.5) the crust appears weaker, with an effective elastic thickness of 80 km. During Sequence B time, however, all models work with an effective elastic thickness of 100 km (Fig. 2.6).

Noticeably, one best-fitting model, line 165_09, shows changes in effective elastic thickness from 80 to 100 km, from one period to the other, while another, line 118_15, keeps the same elastic thickness of 100 km in both periods. Our range of effective elastic thickness (80 to 100 km) falls within the 25% of uncertainty of estimated
accuracy for similar data (Burov and Diament, 1995). These high values probably indicate an unaltered continental lithosphere (no decoupling) after being tectonically loaded (Burov and Diament, 1995). The relatively constant effective elastic thickness through time rules out any significant weakening of the Australian plate, therefore, no relaxation and/or visco-elastic rheology is necessary to model the basin. Moreover, according to our results, loading appears not to have any weakening effect on the elastic thickness of the Australian plate. This is in agreement with the elastic behavior of the lithosphere during foreland time.

Misfits between observations and models, for example in Sequence A lines 116_04 and 118_15 (Figs. 2.5), are probably due to the faulted nature of the foreland sequences. Restoration of faults will not represent the original geometry of the sequence because these faults developed coevally with deflection. The best matches, for example Sequence B, line 118_15, (Fig. 2.6), are obtained in undeformed sequences, where decompacted thickness is a good representation of the geometry of the basin. Results are limited by the length of individual seismic surveys which extend up to 236 km, (line 118_15), whereas deflection extends over 400 km, therefore, the models only match part of the stratigraphic record and uncertainty increases closer to the Trough.

Lateral and spatial variation in crustal strength has been previously reported in foreland basins. For the Bermejo foreland basin in Argentina, Cardozo and Jordan (2001) invoke inherited heterogeneities in pre-bending lithosphere as the cause of these variations. In the Timor Sea, Tandon et al. (2000) develop flexural models for the present day bathymetry and the top of the Pliocene section. The found laterally variable effective elastic thickness (25-75 km), due to changes in curvature and faulting related to inelastic
yielding. Spatial variation in the amount of effective elastic thickness has two probable causes. One could be the effect of an inherited rheologically heterogeneous basement. The central-eastern Timor Sea contains crystalline paleo-highs, which may correspond to regions of higher effective elastic thickness, while the west Timor Sea is affected by ancient basement-grabens filled with pre-Miocene sediments that may decrease lithospheric strength prior to bending (Lavier and Steckler, 1997; Fig. 2.1). The other probable cause could be lateral variation in strain rate or in strain partitioning, along the collision zone (Harris, 1991), resulting in differential bending and models of variable lithospheric strength. The Timor Sea foreland sediments do not reach the 3 to 5 km in thickness that according to Lavier and Steckler (1997) are the minimum values necessary to decrease the effective elastic thickness in areas of crust thinner than 35 km. This agrees with the flexural models of this work that indicate that the foreland sedimentary-cover in Timor Sea appears not to have a significant effect on the effective elastic thickness.

The change in the modeled deflection through time shows that during late Miocene-early Pliocene time, the west Timor Sea undergoes greater deflection than the east Timor Sea. The difference in the point of maximum deflection (position of linear load) is about 1000 m (lines 165_09 vs. Line 118_02). However, sediment thickness along the seismic survey shows that foreland sediments in the east are thicker than in the west. This apparent contradiction is resolved if we consider that the two areas represent different lateral positions within a flexural model. The western part of the survey is located farther away from the linear load, and appears thinner in spite of requiring greater deflection. In contrast, the east Timor Sea sections appear thicker in spite of less flexure at the position of the linear load. During the late Pliocene, the thickness of the
corresponding foreland sequence coincides with the amount of flexure, as the eastern area appears thicker and more deflected than the western part of the survey (Figs. 2.5 and 2.6; Table 2.2). Although in our model the physical causes of loading can be diverse, an end shear vertical force emulates the tectonic (accretionary wedge) and sediment loading. Different shear force values are needed to match the data and are possibly the cause of significant differences in deflection between the eastern and western Timor Sea. It is interesting to note that the amount of deflection (up to ~3500 m) is much larger than the expected subsidence produced by the combination of sea level changes (0-150 m, Haq et al., 1988) and sediment loading. The difference in the modeled deflection between east and west Timor Sea may be due to the polarity of basin closure. Harris (1991) shows an oblique southwestern and northeastern propagation of the collision zone between the Eurasian and Australian plates.

At the end of the early Pliocene, the theoretical flexure involves ~470 km of Australian plate. The point of maximum deflection and the position of the linear load (Po in Fig. 2.9) are located northwestward of the present Timor Island (Fig. 2.9A). The forebulge is between 269 and 310 km wide and ~261 m high. Its top is located in continental Australia, at a point that coincides with today’s westernmost Kimberly Highlands (Fig. 2.9A). By the end of the late Pliocene (~1.6 Ma), the position of the theoretical point of maximum deflection and linear load (Po) had migrated ~120 km southeastward in the west Timor Sea and ~100 km in the central and east Timor Sea (Fig. 2.9B). The remaining elements of the flexural model, such as the flexural node (point of zero deflection, Xo in Fig. 2.9) and the top of the forebulge (Xb in Fig. 2.9) migrated in the same direction. The displacement of the flexural elements (100 to 120 km) is shorter,
although of similar magnitude, than the estimated amount of subducted plate under Eurasia (150 to 200 km). The estimated rates of plate convergence (Tregoning et al., 1994; Genrich et al., 1995) suggest that convergence was not constant during collision. According to our models, at least 570 km of Australian plate have been flexed due to vertical loading.

Interestingly, the position of the youngest landward boundary of the regional deflection (Fig. 2.9B) coincides with the basin-ward boundary of the un-stretched continental crust in the Kimberly Block area (~35 km) as described in the models of Petkovic et al. (2000). This indicates that only stretched continental crust was flexed during collision in the west Timor Sea.

According to our results, the upper lithosphere was experiencing tension followed by normal faulting coeval with thrusting in the accretionary prism and flexure along the whole Australian shelf. The distribution of normal faulting throughout the entire survey and, according to many authors, along the entire Timor Sea (Woods, 1994; O'Brien et al., 1996; Baxter et al., 1998; Lorenzo et al., 1998), precludes any transmission of large horizontal compressional stresses.

Lorenzo et al. (1998) recognize that the present curvature of the Timor Trough ($10^{-7} \text{ m}^{-1}$) appears low compared with values reached by Kruse and Royden (1994) for the Adriatic Sea ($10^{-6} \text{ m}^{-1}$). Our models show an even lower maximum curvature value ($10^{-8} \text{ m}^{-1}$). The difference probably comes from the fact that they analyzed present day bathymetry that represents the added curvature produced during the entire foreland evolution. Whereas we studied two time intervals and assumed they were independent spatial and temporal events.
Figure 2.9. Late Mioce-early Pliocene (A) and late Pliocene (B) geodynamic evolution of Timor Sea. The amount of subducted Australian lithosphere is at least 100 km, according to this model. Effective elastic thickness (EET) is 80 to 100 km for both periods. Flexural variables as defined in Tables 1 and 2. Po: Linear Load, Xo: node point (0 deflection), Xb: position of the top of the forebulge.
According to Watts (2001) and Burov and Diament (1992) curvatures of \( \sim 10^{-6} \text{ m}^{-1} \) may reduce the effective elastic thickness up to 50\%, while values up to \( 2 \times 10^{-7} \text{ m}^{-1} \) may reduce the effective elastic thickness up to 20\%. Lower values do not represent a significant change in effective elastic thickness. Therefore, according to the values determined in the present study for the periods of time represented by sequences A and B, the curvature should not significantly decrease the effective elastic thickness of the Australian plate during continental collision. This agrees with the invariable effective elastic thickness derived from flexural models.

Low curvature values also agree with the small amount of displacement found in flexure-related normal faults. Although inelastic yielding exists, as established by faulting, the results suggest it was not enough to change the regional rheology of the plate. Therefore the elastic rheology is consistent with our models. Interestingly, the position of maximum curvature in the west Timor Sea coincides with the position of the Cartier Trough, using 80 and 100 km of elastic thickness for both periods (Figs. 2.1 and 2.9). In addition to salt withdrawal, part of the Cartier Trough deformation may be attributed to high concentration of strain due to flexure.

Kruse and Royden (1994) invoke dynamic stresses, phase changes and conductive heating as causes of reduction in load in the Adriatic Sea through time. However in the Timor foreland basin, the load reduction estimated along the western area is not easily explained using these causes since it is considered a thermally stable zone. One plausible explanation for the variation in amount of loading in our models is the changing position of the point of maximum bending that migrates towards Australia as the linear load moves in the same direction. The evolution of accretionary prisms conveys continuous
development of new faults and related folds, as well as across-the-strike terminations and relays of these structures. These could all be responsible for the adjustment in the amount of loading affecting the plate through time. Harris (1991) shows the change in geometry of the tectonic wedge during the collision of the Australian and Eurasian plates.

According to our models, the amount of vertical shear stress in both west and east Timor Sea is similar. However, the amount of deformation due to flexure, implicitly inferred from fault displacement, indicates that the west has supported more cumulative strain than the east. The difference in fault displacement is significant during the late Miocene-early Pliocene as the slip is up to 2.5 times greater in the west than in the east. Increasing deformation in the east Timor Sea during late Pliocene time probably indicates an increase in the amount of stress propagation in the same direction. This polarity of fault activity in the Australian plate corroborates the oblique nature of plate collision suggested by our models of plate deflection through time and supported by previous workers (Hamilton, 1979; Harris, 1991).

Summary

Flexural models show that during foreland basin evolution (~6.5 to 1.6 Ma.), the effective elastic thickness of the Australian lithosphere in the Timor Sea is between 80 and 100 km. These elastic thickness values agree with the old geothermal age of the Australian Plate. Spatial changes in plate strength are due to heterogeneities prior to bending or basement distribution. The east Timor Sea displays thicker crystalline basement (paleohighs) and appears to be stronger than the west Timor Sea. The latter exhibits thicker pre-bending sedimentary-cover filling wide graben structures. However, the difference in effective elastic thickness between both areas falls within the estimated
error. The effective elastic thickness does not appear to change during foreland basin evolution. Consequently, no visco-elastic relaxation can be inferred from the flexure of the plate during loading. Foreland sediment cover is too thin to cause any lithospheric weakening. Curvature of the Australian plate in the Timor Sea is low when compared to other foreland basins. The small curvature was not enough to weaken the lithosphere in the area. Accordingly, normal faulting exhibits small apparent slip (up to 150 m). Moreover, inelastic yielding appears not to affect the regional rheology of the Australian lithosphere.

According to our models, at least 570 km of the Australian plate (mostly areas of stretched continental crust) was flexed due to vertical loading during the collision. However, the deflection at one particular time involved only 400 to 470 km (most of the Australian Platform). As the tectonic loading advanced towards continental Australia (100 to 110 km), an increasing portion of the Australian plate became flexed. These figures correspond to the estimated value of subducted plate. According to our models, the point of maximum deflection was located under today’s accretionary prism during the entire period of foreland evolution. The top of the forebulge was located in continental Australia, and is partially represented by the Kimberly Highlands.

The deflection of the plate caused by tectonic loading was at a maximum in the west Timor Sea (up to 3500 m) at the beginning of the collision and at a maximum in the east Timor Sea (up to 2300 m) by the end of the convergence. This eastward propagation of the deflection indicates the oblique polarity of the collision between the Eurasian and Australian plates.
Flexure-related normal faulting corroborates the oblique character of the collision. Faulting begins to affect the plate in the west Timor Sea at the beginning of convergence and propagates eastward in the same direction of the deflection. The west Timor Sea exhibits higher accumulated deformation due to a longer time of loading and coeval bending of the plate.

References


CHAPTER THREE. FLEXURE CONTROL OF ACCOMMODATION IN A MODERN FORELAND BASIN: THE NW SHELF OF AUSTRALIA, TIMOR SEA

Introduction

Foreland basins are asymmetrical depocenters, whose subsidence is controlled, primarily, by the reaction of the elastic lithosphere to the combined effects of downward forces (e.g., sea water, tectonic, and sedimentary loads), and upward forces (e.g. mantle buoyancy) (Beamont, 1981; Jordan, 1981; Puigdefabregas et al., 1986; Heller et al., 1988; Flemings and Jordan, 1989; Coakley, 1990; Shanley et al., 1994; Jordan, 1995; Cardozo and Jordan, 2001). Although these lithospheric adjustments modify the regional basin geometry and sedimentary patterns, they have been largely ignored in foreland basin stratigraphic studies, which in stead rely mostly on eustatic changes (e.g., Cretaceous Interior Sea belt in the continental USA). In this study, the geometry of the stratigraphic units that fill the resultant accommodation are forward modeled at different times during the evolution of the basin to evaluate the amount of subsidence caused by sediment and tectonic loading throughout its history. This evaluation can also reveal changes in the effective elastic thickness of the lithosphere, the regional geometry of the first order deflection and highlight their interaction with stratigraphic development.

The response of the elastic lithosphere to forces depends on the model assumed. Local isostatic compensation, for example, requires the rigidity of the lithosphere to be zero (Turcotte and Schubert, 1982; Allen and Allen, 1990; Watts, 2001). Local compensation can be assumed for eustatic fluctuations (Bloom, 1967), since they are considered to be of global nature and cover an entire plate. Watts (2001), however, shows that at time scales of $10^4$-10$^7$ m.y., flexural compensation (regional isostatic
compensation) is in general accord with most of the remaining geological observations. Therefore, foreland basins are often modeled as a regional isostatic compensation to tectonic-thrust belt- loads (Jordan, 1981; Quinlan and Beamount, 1984) and sediment loads (Sclater and Christie, 1980; Coakley, 1991; Cardozo and Jordan, 2001).

Foreland basins are produced at convergent-plates margins, adjacent to orogenic belts. Figure 3.1 summarizes all possible forces that act on convergent-plate settings. In order to study the resultant subsidence or uplift of the plate at one instant of time, it is assumed that the acting forces are in equilibrium. Which specific forces dominate depends on the tectonic setting of the basin and age of basin formation. Thermal-related subsidence, for example, may contribute significant portion of the total subsidence up to tens of million of years after a rifting event (McKenzie, 1978; Sclater et al., 1980). Slab pull has been associated with the weight of an oceanic crustal slab that is descending into the upper part of the mantle, pulling the plate toward the subduction zone (Royden, 1993). Mitrovica et al. (1989) explain the >1500 km wide Alberta basin deflection by invoking secondary mantle convection (dynamic loading, Fig. 3.1). The restoration force of the mantle tries to reinstate the plate to its initial position (Turcotte and Schubert, 1982). The rigidity of the plate, expressed as effective elastic thickness (EET) in kilometers, is the measure of the integrated brittle, elastic and ductile strength of the lithosphere (Watts, 1978; Turcotte and Schubert, 1982, Watts and Burov, 2003). Different processes can produce changes in the rigidity of the plate. According to Lavier and Steckler (1997) sedimentary covers can thermally isolate the lithosphere and induce changes in plate strength (EET). Curvatures of the plate in excess of $10^{-7}/m$ might induce inelastic yielding, thus, the lithosphere will appear weaker than it actually is (Burov and
Diament, 1992; Watts, 2001). Lateral changes in crustal composition might result also in effective elastic thickness differences (Stewart and Watts, 1997). In addition, plate strength increases with time in thermally young cooling lithospheres (Bodine et al., 1981). Watts (2001), on the other hand, predicts EET changes with time because ductile deformation is a function of strain rate. With passage of time the strain-rate decreases, as does the depth of the onset of ductility. This results in a thinning of the elastic core of the lithosphere, reducing its strength.

Figure 3.1 Conceptual tectonic setting for foreland basin formation (Modified from Catuneanu et al., 1997)
The stratigraphy of the basin is the best measure of the development of the vertical movements of the plate through time. Any sedimentary column implicitly represents the minimum resultant accommodation created for self-content (sensu Jervey, 1988). The total space available for accumulation of sediments (accommodation) is the result of the interplay between subsidence and base-level (eustasy). According to Emery and Myers (1996) the change in accommodation per unit of time can be expressed as:

\[ \Delta \text{Accommodation} = \Delta \text{base level (eustasy)} + \Delta \text{subsidence} + \Delta \text{compaction}. \]

Where \( \Delta \) means change per unit of time. For a given sedimentary package, its decompacted thickness represents the minimum accommodation available at depositional time.

In addition, according to Price (1973), there is a link between foreland stratigraphy and thrust-belt development. Through this association, the sedimentary succession of the basin records the history of the thrust wedge, since tectonic pulses not only create accommodation for sedimentation but also are an important source of sediments. Beamont (1981), Jordan (1981), and Flemings and Jordan (1989) have tried to reproduce this relationship using flexural and depositional mathematical models.

Using seismostratigraphic information as a primary data-base and forward modeling on decompacted sedimentary-thickness data from the well-documented late Tertiary Timor Sea peripheral foreland basin (northwestern Australia), this study tries to assess the effects of the tectonic and sediment induced plate-deflection on regional stratigraphy at a third order marine-cycle scale (sensu Van Wagoner et al., 1990) and whether the lithospheric-strength (EET) changes during basin history. Input data for
modeling, such as the geometry of sedimentary loading and thrust-belts, are taken from geological observations in the studied area.

**The Flexural Model**

Oceanic and continental flexural studies suggest that the long-term behavior of the lithosphere can be modeled, to a first order, by the deflection of a thin elastic infinite or semi-infinite plate (the lithosphere), over an inviscid fluid (Turcotte and Schubert, 1982; Fig. 3.2, Appendix 1). The inviscid fluid acts with a force that at every point is proportional and instantaneous to the deflection (Hetenyi, 1946). This is equivalent to the restoration force of the mantle that tries to return the deflected lithosphere to its original configuration (Turcotte and Schubert, 1982). The estimated effective elastic thickness of the plate can vary both in time and space and has provided information about the relationship between the loads, the age and some mechanical properties of the plate (Watts, 2001). Tectonic and sedimentary loads are viewed as producing independently distributed vertical stresses acting on discrete parts of the elastic plate at different times. The total contribution to deflection for each load is calculated from the resultant sum of individual smaller loads, integrated using the superposition principle (Hetenyi, 1946, Appendix 1). The deflection produced by each load differs from the other, therefore the resultant regional flexure is the summation of all deflections (a complete explanation of flexure theory can be found in Appendix 1).

Blair and McPherson (19941) argue that tectonic uplift rates are on average eight times higher than average denudation rates, thus, a time lag exists between uplift and sedimentation. Therefore, for the purpose of the calculations, the flexural evolution of a foreland basin during a tectono-stratigraphic event is assumed to be divided into 2 steps:
1) tectonic loading and instantaneous deflection of the lithosphere, and 2) sedimentation and deflection caused by sedimentary load which is evaluated at the end of the sedimentation period (Fig. 3.3). Vertical movements originated in response to local isostatic compensations caused by eustatic changes are introduced when necessary. Regional sedimentary cycles, punctuated by basin-wide events such as unconformities, maximum flooding surfaces and paleontological occurrences, are used as spatial and temporal units for flexural analysis.

Figure 3.2. Physical model of flexure for elastic beams. A) Deflection due to the a linear load with different flexural rigidities (effective elastic thickness) (after Beamout, 1981). B) Changes in flexure due to changes in size and distribution of the vertical load (after Watts, 2001).
The Effective Elastic Thickness

The concept of effective elastic thickness in continental lithosphere is still controversial because, unlike oceanic lithosphere, it can not be related to an isotherm or thermal age (Burov and Diament, 1995; Watts, 2001). The effective elastic thickness reflects the integrated brittle, elastic and ductile strength of the lithosphere on time scale > $10^5$ years (Watts and Burov, 2003). As part of the lithosphere, continental crust has usually a more complex evolution and is potentially more heterogeneous than oceanic crust because it is on average less dense and much older than oceanic crust, therefore it does not subduct easily, and has more chances of being incorporated into orogenic belts or undergo stretching during rifting and closing of oceans, as well as of being intruded.
(Dewey and Bird, 1970). Thus, the EET estimates in continental plates may be highly biased toward inherited mechanical properties from previous rifting or orogenic events (Beamout, 1981; Burov and Diament, 1992; Kruse and Royden, 1994; Cardozo and Jordan, 2001).

Rheologic models of the lithosphere suggest that the strength of the lithosphere is limited by brittle and ductile deformation (Burov and Diament, 1992; Kruse and Royden, 1994; Lorenzo et al., 1998). If an applied load exceeds the strength of the plate, it yields rather than flexes and the plate will appear weaker than it actually is (Lorenzo et al., 1998; Watts, 2001). As a result of these models it has been established that yielding will change EET for a particular curvature (Goetze and Evans, 1979; McNutt, 1984). Curvatures greater than $10^{-7}$ m$^{-1}$ in elastic plates, will decrease the EET because of yielding in the upper and lower part of the plate. If the curvature is high enough, the plate may fail completely (Watts, 2001). Similarly, if an elastic-plastic rheology is assumed, the plate will bend elastically until stress exceeds the yield strength. Zoetemeijer et al. (1999) establish the boundary fiber stress for the onset of plastic yielding at about 1 GPa.

**Vertical Loads**

The geometry of modern foreland basins is the result of changes in size and shape of vertical loads over time (Coakley, 1991; Watts, 2001). Figure 3.2 shows how the lithosphere responds to different geometries of vertical loads. Distributed loads (Fig. 3.2B) tend to produce wider basins than linear loads. In this study, we implement distributed loads. We consider that they are a better approximation to the nature of foreland basins. Concentration of loading at the end of the plate leads to overestimation
of the amplitude and underestimation of the maximum-width of the first order deflection. These differences are accentuated when a weaker plate is used.

**Eustasy**

Eustasy refers to global variation in sea level referenced to the center of the earth caused by changes in the volume of ocean-water or in changes in the size of the ocean-basin (Emery and Myers, 1996). A rise in sea level causes homogeneous subsidence in the ocean floor due to the additional weight of the newly added water column. Inversely, isostatic rebound occurs when sea level drops (Fig. 3.4). Allen and Allen (1990) estimate that sea floor isostatic subsidence/rebound is about 0.4 of the eustatic fluctuation. In other words, a eustatic sea level rise of 1 m generates 1.4 m of additional accommodation (water depth) (assuming a mantle density of 3300 kg/m$^3$).

This study models the accommodation rather than subsidence. Therefore, for the purpose of the accommodation equation, the amount of sea level rise is added to its local isostatic compensation and the amount of sea level drops subtracted from its local isostatic rebound. The global sea level curve produced by Haq et al. (1988) is used here as input data. The accuracy of the global chart has been the subject of many debates (Watts, 1978; Miall 1997; Watts, 2001). Many authors have chosen to assume a constant sea level for modeling purposes (Coakley, 1991; Jordan and Flemings, 1991). For this work, the cycle chart is not taken as an accurate value, but rather as a way to explore the effect of global sea level fluctuations in the model, when a reference datum is used.
Lithospheric compensation of the weight of sedimentary columns was recognized in early studies of basin analysis (Holmes, 1968). One-dimensional backstripping techniques apply this concept to determine the burial history of basins (Sclater and Christie, 1980). However, they assume local isostasy. The advent of flexural models in basin analysis leads to the assumption that sedimentary loads form part of an integral linear load at the end of the plate (Turcotte and Schubert, 1982; Beamont, 1981; Jordan, 1981). More recent studies use distributed loads of predetermined geometry to model the
flexure of the basin caused by sediments. Allen and Allen (1990) and Roberts et al., (1998), following the sinusoidal topographic load-geometry of Turcotte and Schubert (1982), model the flexural deflection of sediments in passive margins according to a compensation factor (ratio between regional and local compensation) and the wavelength of the basin. Peper (1994), Lorenzo et al. (1998), and Sarmiento (2002) use triangular and rectangular geometries to represent the different amounts of sediment loading.

The model described herein uses a sedimentary geometry taken from the seismostratigraphic record, for example sedimentary wedges (Fig. 3.5). Values of decompacted thickness, density, length, and the angle of the wedge (slope of the shelf/ramp) are estimated for each analyzed sedimentary unit. Appendices 2 and 3 show in more detail the decompaction techniques used to carry out these calculations. Interval velocities are estimated from stacking velocities of the survey. They range between 1500 to 2500 m/s (TWTT). The resulting error in this thickness calculation might be 11 to 14%. From the thickness error introduced by interval velocities inaccuracy, porosity can change up to ~3%. This in turn introduces an error in decompacted thickness up to ~7 %, which we consider an acceptable value since the average thickness change from compacted to decompacted thickness is ~28%. The extent of the sedimentary load comes from geologic records at each particular modeled-time. The results should be reasonably consistent with the entire evolution of the basin, i.e. with previous and subsequent sediment distributions.

**Tectonic Loading**

Thrust and fold belts produce enough crustal thickening to generate lithospheric flexural compensation (Beamount, 1981; Jordan, 1981). For this reason, they are thought
to be the primary cause of depocenter development in foreland basins (Beamount, 1981; Jordan, 1995; Fig. 3.1). Geodynamic studies in foreland basins use diverse geometries to represent the shape of tectonic loads. Lorenzo et al., (1998), for example, use basin floor topography and Wetar Island topography and triangular geometries to represent supracrustal loads along the Timor Trough. Coakley (1991) maintains a constant upper part of the thrust-wedge slope (via erosion), throughout its permanent progress to estimate the amount of loading at any particular time. Most of them, however, do not constraint the shape of these loads with geological observations and keep the geometry constant despite changes in flexure related to their geometry (Fig. 3.3).

For the purpose of modeling the deflection induced by tectonic loading, it is assumed that the thrust-belts behave as a Coulomb material (Davis et al, 1983; Appendix 3). Thus, the amount of loading is distributed along the tectonic-wedge, whose length and thickness is a function of its taper angle that remains constant during the entire evolution of the wedge. In this study, we try to constrain the geometry and taper angle of the tectonic load (thrust belt) from local cross-sections, consistent with the geology of the analyzed area. In addition, results can be interpreted in terms of thrust belt dynamics since the length of the wedge, measured from the hinterland toward the foreland, represents the total width of the active belt and therefore may represent the position of the tip of the wedge or deformational front.

**Implementation of Flexural Modeling**

Implementation of flexural modeling begins with the decompaction of the modern thickness of each sedimentary cycle taken from seismostratigraphic data as a function of present day-porosity (Fig. 3.5, Appendix 3). It is assumed that the top of the decompacted
Figure 3.5. Flexural modeling: 1) Seismic interpretation of marine cycles. 2) Decompaction of sedimentary load and discretization (Appendix 2). 3) Flexural modeling and correction by eustasy. 4) From decompacted thickness, flexure due to sediment and eustasy is subtracted, thus, the tectonic induced subsidence is obtained. 5) Flexural modeling of tectonic induced deflection.
cycle was at base-level and its thickness represents the accommodation of the basin at depositional time after tectonic and sedimentary loads have deflected the lithosphere. This assumption is necessary, because water depth data is not available. Thus, onlap in seismic data is assumed to represent local sea level changes that are used as the upper boundary of accommodation.

The amount of flexural subsidence caused by sediment loading is then calculated. The sediment load is discretized in individual linear loads whose individual deflection is later integrated using the superposition principle (Hetenyi, 1946; Appendix 1). The amount of deflection obtained, is adjusted by the compensation caused by the eustatic fluctuation during the elapsed time represented by the cycle.

The results from this calculation are subtracted from the original decompacted thickness. If the difference between these values is negative, it is assumed that no tectonic loading occurred during the sedimentary cycle. However, if the difference is positive, it is taken to represent the minimum accommodation created by thrust-wedge loads. This difference is forward-modeled as a function of the amount of tectonic loading and effective elastic thickness until a subjective best-match is obtained. This sediment-thickness “excess” is assumed to represent the minimum water depth profile at the time just before deposition. Unconformities are not taken into account since uncertainties about the erosion rate, missing time and the diachronous character of erosional surfaces are too large to be satisfactory quantified.

The computer routine used here was developed using MatLab 6.5® (Appendix 4) and it is devoted entirely to flexural modeling. Therefore, the decompaction calculation is not included. The routine was tested using different parameters such as linear loads, and
compared with distributed load programs implemented by Cardozo and Jordan (2001). Geodynamic constants such as the Poisson’s ratio, Young’s modulus, gravity acceleration and densities (water, mantle, decompacted sediments and thrust-wedge) are taken from published data. The sediment load is divided into discrete 10 km-wide rectangles, whose heights are a function of the decompacted thickness change along the seismic profile (Appendix 2). The difference in resultant sedimentary-deflection using 1 km wide and 10 km wide loads is between 2 to 3%. For computational efficiency, we implemented wider loads. Similarly, the tectonic wedge is divided into discrete, 1 km-long rectangles whose heights are a function of the critical taper angle (Appendix 2). The output shows the deflection caused by each load and the summation of both deflections.

**Case Study: The Tertiary Foreland Basin in Northwestern Australia, Timor Sea**

The late Tertiary foreland basin in the Timor Sea provides an excellent example of a peripheral foreland basin (Jordan, 1995; Lorenzo at al. 1998; Tandon et al., 2000; Fig. 2.1). It developed over attenuated Australian continental crust during the latest collision episode between the Australian and Eurasian plates, approximately 6.5 to 1.6 Ma (Hamilton, 1979; Fig. 3.6). In this area, the stretching factor ($\beta$) calculated in Precambrian basement rocks is 2.6 (Petkovic et al., 2000) and the thermal age of the pre-Tertiary basement is early Jurassic (Baxter et al 1998). Therefore, the subsidence related to thermal cooling during the foreland period is negligible. The Timor Trough, a deep depression, is created by the deflection of the Australian plate under the load of the accretionary wedge (Hamilton, 1979; Harris, 1991), as well as sediment and sub-surface loading (Lorenzo et al., 1998).
Since the late Miocene, the North West Australian Shelf in the Timor Sea has been a carbonate ramp, whose stratigraphic architecture has been primarily driven by local sea level fluctuations (Apthorpe, 1988). The lithology of the entire succession is described in some wells as greenish-gray calcareous claystones and silt stones of outer shelf, shelf edge and platform environments (Apthorpe, 1988). The carbonate succession is underlined by a regional unconformity that represents the beginning of the collision between both plates in the area, and has been identified in the commercial wells.

Figure 3.6. Crust distribution in Timor Sea area. Note the extent of the continental crust beyond Timor Island (Hamilton, 1979).
Delambre 1 and Buffon 1 (Fig. 3.1; Aptorphe, 1988). This regional unconformity, found at the top of Middle Miocene limestones (planktonic forams of Zone N10), is overlain by late Miocene carbonate sediments of deep-water nature (Zone N15) (Aptorphe, 1988). A second unconformity, reported by Veevers et al. (1974) at the ODP Leg-site 262, separates shallow water upper Pliocene dolostones and calcarenites from folded shallow-water lower Pliocene carbonates at the axis of the Trough. Boehme (1996) also identifies this unconformity along proprietary seismic profiles. According to Hillis (1992), this hiatus represents a short-term depositional break (< 1 My) at Ashmore Reef 1 well (Fig. 1.1). The sedimentary succession overlying the lower Pliocene unconformity is punctuated by the base of unconsolidated sediments (Apthorpe, 1988). Hamilton (1979) and Lorenzo et al. (1998) show seismic lines along the Timor Trough in Australia with significant reduction in sedimentation toward the walls of the Trough.

**Data**

Over 2000 km of two-dimensional seismic reflection data (4 to 6 seconds two-way travel-time records) from the Australian Geological Survey Organisation (AGSO) 1996 seismic program serve as primary source data for the study of the modern Australian platform and part of the outer wall of the Timor Trough (Fig. 2.1). Well-log information and seismic data from the literature are used to tie biostratigraphic and lithological data to the seismic record (Apthorpe, 1988; Wormald, 1988; Ostby and Johnstone, 1994; Woods, 1994). Two seismic lines are chosen as representative of the foreland basin: line 165-9 in the southwestern area (Fig. 3.7) and line 118-15 in the northeastern area (Fig. 3.8). They show the foreland sequence in an almost pristine stage, with no significant deformation.
**Seismic Interpretation**

Seismostratigraphic interpretation of foreland sequences is carried out using a classic sequence stratigraphy approach (i.e., assuming that the regional reflector geometry is developed primarily from changes in sea level, Vail et al., 1977). Marine cycles (sedimentary cycles) are identified using onlap shifting between reflectors. The cycle is bounded at the top by the landwardmost shift in onlap of the reflector in one cycle and at the base for the hinterlandmost shift in onlap of the reflector overlying the previous cycle (an onlap shift from landward to hinterland should be identified; Figs. 3.7 and 3.8). The rate of sedimentation is estimated between ~150 to 180 m/m.y. (Appendix 3). These rates were used to determine the age of reflectors between the three dated regional unconformities (Figs. 3.7 and 3.8).

Seven marine transgressive-regressive cycles are interpreted along line 165-9 (approximately 131 km long, Fig. 3.7). The lower late Miocene to early Pliocene sequences (4 in Fig. 3.7) increase in thickness toward the Trough, whereas the late Pliocene sequence thickening is less marked toward the Trough (3 in Fig. 3.7). Along line 118-15 (approximately 236 km long, Fig. 3.8) only 5 sequences are interpreted in the entire foreland succession and all of them show clear thickening toward the trench. It is interesting to note the difference in the number of sequences between the east and west Timor Sea during a given period of time. In the late Miocene to early Pliocene time, four marine transgressive-regressive cycles are identified in the west (Fig. 3.7), whereas only two are identified in the east (Fig. 3.8). It becomes difficult to invoke eustatic changes to account for this disparity, since in classical sequence models sea level changes affect coevally the entire basin.
Figure 3.7. Seismic data and interpretation of line 165-9 (see Figure 2.1 for location). Note the clear wedge-like geometry on the foreland sequence outlines in black lines. Seven marine cycles are interpreted in this data. Age is based on published data (Veevers et al., 1974; Apthorpe, 1988; Woods, 1994; Boheme, 1996) for 6.5, 3.4 and 1.6 Ma. reflectors) and in sedimentation rate calculation (Appendix 3).
Figure 3.8. Seismic data and interpretation of line 118-15 (see Figure 2.1 for location). Compare the number of sequence with line 165-9. In this line only 5 marine cycles are interpreted. In particular between 6.5 to 3.4 Ma., only two sequences are identified, in contrast with 4 marine sequences found along the line 165-9 during the same period. Age is based on published data (6.5 and 3.4 Ma. reflectors) and in sedimentation rate calculation (Appendix 3).
Seismostratigraphic Data Reduction

The simplest approximation to the regional geometry of the basin is the wedge-like shape of the decompacted stratigraphic record at each marine cycle. The thickness of the marine cycles is calculated from interval velocities (2000 to 2500 m/s, TWTT). These values are estimated from the stacking velocity following Telford et al. (1990) and corrected with well information (Ashmore Reef-1; Delambre-1; Buffon-1; Fig. 2.1). Decompaction of the sediment packages to initial conditions for each cycle is carried out according to Sclater and Christie (1980) and Watts (2001) (Appendix 3). These conditions, include an original porosity of 63% and a porosity exponential constant estimated in 0.57/km (Hillis, 1990). The grain density is calculated as 2700 kg/m³ and the density of infill sediments is estimated as 1800 kg/m³ (Appendix 3). The geometry of the sedimentary load is determined using the present-day ratio between the thickness at the edge of the platform (Line 165-9) and the thickness at the toe of the thrust-front, near the axis of the Trough (Veevers et al., 1974, ODP 262). ODP Leg 262 shows ~340 meters of Pleistocene/Holocene sediments deposited at the toe of the thrust-wedge. This ratio is estimated to be 1:3 and it is assumed is the same for all marine cycles.

The geodynamic constants used during modeling are shown in Table 3.1. These values remained constant during the entire modeling process. The density of decompacted sediment is calculated using published data for sediment-grain and water densities (Hillis, 1990).

The critical taper angle and density has to be estimated from publications showing sections across the Timor Island accretionary prism (Table 3.2). A critical taper angle of 6° is used for modeling the tectonic wedge in Timor Sea (Appendix 2). This was the
approximate average of sections based on seismic data. This value lies within the range of taper angles of global thrust belts of 1° to 12° (Davis et al., 1983).

Table 3.1 Geodynamic constants used during modeling.

<table>
<thead>
<tr>
<th>Constant</th>
<th>Symbol</th>
<th>Value</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gravity acceleration</td>
<td>g</td>
<td>9.8</td>
<td>m/s²</td>
</tr>
<tr>
<td>Water density</td>
<td>ρw</td>
<td>1035</td>
<td>kg/m³</td>
</tr>
<tr>
<td>sediment density</td>
<td>ρs</td>
<td>1800</td>
<td>kg/m³</td>
</tr>
<tr>
<td>Density of tectonic wedge</td>
<td>ρsw</td>
<td>2300</td>
<td>kg/m³</td>
</tr>
<tr>
<td>Density mantle</td>
<td>ρm</td>
<td>3300</td>
<td>kg/m³</td>
</tr>
<tr>
<td>Young Modulus</td>
<td>E</td>
<td>5e11</td>
<td>Pa (kg/m·s²)</td>
</tr>
<tr>
<td>Poisson’s ratio</td>
<td>ν</td>
<td>0.25</td>
<td></td>
</tr>
</tbody>
</table>

Table 3.2. Thrust-wedge taper-angle data and sources for Timor accretionary prism.

<table>
<thead>
<tr>
<th>Author</th>
<th>Year</th>
<th>Taper Angle (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hamilton</td>
<td>1979</td>
<td>6</td>
</tr>
<tr>
<td>Bowin et al.</td>
<td>1980</td>
<td>7</td>
</tr>
<tr>
<td>Johnston and Bowin</td>
<td>1981</td>
<td>7</td>
</tr>
<tr>
<td>Audley-Charles</td>
<td>1986</td>
<td>8</td>
</tr>
<tr>
<td>Charlton</td>
<td>1989</td>
<td>5</td>
</tr>
<tr>
<td>Harris</td>
<td>1991</td>
<td>10</td>
</tr>
<tr>
<td>Jordan</td>
<td>1995</td>
<td>4</td>
</tr>
<tr>
<td>Harris et al.</td>
<td>2000</td>
<td>5</td>
</tr>
<tr>
<td>Petkovic et al.</td>
<td>2000</td>
<td>9</td>
</tr>
</tbody>
</table>
**Results**

The effective elastic thickness of the Australian lithosphere in the Timor Sea is estimated to be between 60 and 100 km, with the best matching between observation and models at values between 70 and 80 km (Figs. 3.9 and 3.10). This range is broader than the one calculated in Chapter 1 using a linear load without eustatic variation (80 to 100 km). However, it differs only slightly from estimates reported in the literature for this area: 25-75 km (Lorenzo et al., 1998); 90-130 km (Zuber et al., 1989), and 100 km ± 30km (Fairhead, 1997). The rigidity of the plate appears not to change during basin evolution. Match between observations and models in some cycles are difficult to obtain, particularly in areas where sedimentary thickness reduces to zero. In these areas emphases in matching observations and models was primarily along the first-order deflection.

At least 500 km of the Australian lithosphere have been deflected since the late Miocene as a consequence of the collision between the Eurasian and Australian plates in the Timor Sea (maximum width of the first-order deflection). The flexure history of the Australian lithosphere in the west Timor Sea shows a higher deflection early in the evolution of the basin than during its final stages (seismic line 165-9, Fig. 3.9). Late Miocene to early Pliocene cycles show a maximum accumulated deflection of 1200 to 1600 m. Late-Pliocene deflection decreases to only 650 to 700 m. By contrast, in the east Timor Sea, flexural analysis reveals an increase in flexure from ~ 600 m during the late Miocene-early Pliocene to 800 m during the late Pliocene (Lines 118-15, Fig. 3.10). As a consequence, a clear asymmetric deflection affects the Australian plate in the Timor Sea by the end of the early Pliocene (~3.4 Ma), with the western part of the plate being deflected up to 100% more than the eastern part (Fig. 3.11). Maximum deflection rate
varies from ~ 350 ± 50 m/m.y. in the west to 200 ± 20 m/m.y. in the east. Despite the asymmetrical deflection, 2-D flexural analysis is still valid because the first order deflection occurs primarily along the regional stress field, parallel to the tectonic transport and perpendicular to the collision zone. Differential deflection along the strike is originated by a mechanical respond to the diachronous character of the developing thrust belt and consequent differences in the amount of loading. Wessel (1996) argues that this results in an overestimation of the amount of loading or underestimation of effective elastic thickness in more deflected areas, and conversely, underestimation of the amount of loading and overestimation of the effective elastic thickness in less deflected areas. In addition, the same author notes that the standard over/under estimation in 2-D models is around 20%, which is smaller than uncertainties associated with the nature of the rheology of the lithosphere. 3-D modeling generates uncertainties smaller than those of lithospheric rheology (about 25%) (Ibid.). It also requires accurate control on the geometry of loads to have significane, which in the case of Timor Sea accretionary prism is hard to obtain at the time.

By the end of the late Pliocene, a change in deflection polarity occurs as the plate in the east is deflected ~16% more than in the west. The highest curvature calculated in one individual cycle (Line 165-9, 0-02) is 5x10^{-8} m, at the maximum bending moment (2.4x10^{17} Pa), about 160 km south of the thrust belt. Fiber stress calculated for 100 km of EET is on the order of 1.3 ± 0.26 GPa and for 60 km is 0.8 ± 0.16 GPa. These values are around the onset value for plastic yielding (1GPa). They also agree with the small-slip normal faulting found in the foreland. The plate was on the verge of plastic yielding when the flexure process ceased.
Figure 3.9. Flexural modeling results for sequences in line 165-9 (southwest Timor Sea). Best matches are found with effective elastic thickness between 70 and 100 km. Dashed lines represent decompacted sediment thickness corrected by eustasy where it applies. Geometry of distributed loads is shown in Figure 3.12.
Fig. 3.10. Flexural modeling results for sequences in line 118-15 (northeast Timor Sea). Best matches are found with effective elastic thickness between 70 and 100 km. Dashed lines represent decompacted sediment thickness corrected by eustasy where it applies. Geometry of distributed loads is shown in Figure 3.13.
Figures 3.12 and 3.13 show the total deflection and the contribution of each component (sediments, eustasy and tectonic load) to each marine cycle, as estimated using solely seismostratigraphic and well data. During late Miocene-early Pliocene, tectonic loading recorded in accommodation was responsible for at least 30 to 47% of total deflection, compensation due to sediments accounted for an additional 50 to 63% and eustasy for the remaining 3 to 7%. The highest period of tectonic loading was reached close to the 3.4 Ma unconformity. During late Pliocene the tectonic flexure was
minimal (18 to 21%) and sediment-compensation accounted for the remaining 78 to 81%. Today, however, the Timor Trough is covered with a water-column of up to 2500 m that are not reproduced in our models because we do not have water depth data. Therefore, the “excess of deflection” represented by today’s water column, must be added to the total maximum tectonic subsidence. From the total ~5000 m of maximum basement deflection during foreland basin evolution, ~3500 m, i.e. 70%, must be assumed to be due to tectonic loading. Therefore, the effect of eustasy is reduced to about 1 to 2% of the total and sediments must be responsible for the remaining ~28%.

In terms of plate-deflection geometry, sediment-related flexure is 150 to 200 km wider than the one produced by tectonic loading in one single cycle (i.e. approximately 35% wider). The forebulge produced by tectonic loading is higher and narrower than the forebulge produced by sediment loading. Its position is closer to the thrust belt, located in the wall of the trough. Later, these forebulges are depressed by the deflection induced by sediments. In addition, the first order deflection of the sedimentary-flexure produces a very subtle forebulge. These wide structures (~ 100 to 150 km in Timor Sea) overlap and depress those previously produced by tectonic and sedimentary loading, resulting in even shallower forebulges. No clear forebulge is discernible along the seismic sections in Timor Sea. The composite forebulge (Figs. 3.12 and 3.13) is located beyond the data window, closer to the Australian coast.

The amount of loading produced by sedimentary wedges is of the same order of magnitude of the one produced by the minimum acceptable tectonic loads ($10^{10}$ to $10^{11}$ N/m). The distributed sedimentary load is between 350 and 510 km wide. The tectonic load causing the deflection is between 9 and 26 km long. Note that these values of
loading should be taken as minimum because they represent the subsidence caused by tectonic loads recorded in the sedimentary record. Tectonic subsidence, however, might occur without sedimentation.

**Discussion**

Model results are limited by the parameterization of the loads. Sedimentary loads, for example, are extrapolated using modern-day observations. However, a change of 25% km in the width of the load (which is about the uncertainty we have) will produce a change of 26% in the sedimentary deflection that might represent between 20 and 30% in the total deflection. Changes in the slope of the sedimentary basement introduce dramatic changes in the amplitude of the deflection. A regional change of 1 degree introduces changes of two-orders of magnitude in the deflection. A more realistic regional change of $0.2^\circ$, for example, introduces changes of about 11% in the deflection. These large ranges are due to the width of sedimentary covers that at one single cycle can reach 500 km. Tectonic loads are represented as residual flexure in the models. The geometry of these loads is small compared to the length of the plate. Thus, a change of $1^\circ$ in the critical taper-angle introduces a change of ~10% in the tectonic deflection, and depending upon the sedimentary load geometry, the change in total deflection can be negligible.

Since water depth is unknown, we calculated the compensation of sediment as if they would were deposited initially along a horizontal surface and corrected by eustasy. It implies that topography would merge (Figs. 3.12 and 3.13), and explain the large amount of sediment related- subsidence when only sediments are considered. However, the present water depth in Timor Sea indicates the strong tectonic-induced component in the deflection history of the basin. The tectonic subsidence recorded in the stratigraphic
record is between 500 and 700 m, which is only about 25% of today’s water depth at the
deepest part of the Trough. The remaining ~75% is not in the stratigraphic record and the
temporal distribution is uncertain, this is a limitation of the technique used here, and
therefore the obtained values represent the minimum tectonic component for each cycle.

The effective elastic thickness calculated in this study for the Timor Sea foreland
basin partially agrees with previous estimates of the present effective elastic thickness of
the lithosphere in Timor Sea based on forward modeling (Tandon et al., 2000; Lorenzo et
al., 1998) and agrees thoroughly with values obtained by spectral analysis (Zuber, 1989;
McKenzie and Fairhead, 1997). The changes in effective elastic thickness estimated
during the evolution of the Timor Sea foreland basin (i.e., 70 to 100 km) fall within the
25% estimated error associated with plate rigidity studies (Burov and Diament, 1995).
Therefore, forward modeling reveals no significant changes in plate rigidity over time.
The low curvatures obtained by modeling at individual marine cycles agree with the
constant strength of the plate, as values lower than ~ 10^{-7} m are not enough to weaken it
(Watts, 2001), though, as discussed in Chapter one Timor Trough lithosphere appears to
be at the edge of regional inelastic yielding. The sedimentary cover is not thick enough to
cause thermal adjustments. The values of effective elastic thickness obtained in this work
are indicative of a very strong Australian plate.

Line 165-09 in the west Timor Sea exhibits 4 marine cycles occurring during the
late Miocene to early Pliocene, whereas Line 118-15 in the east exhibits only 2. Eustasy
cannot be responsible for this disparity. The differential subsidence produced by the
resultant tectonic asymmetrical deflection (from up to ~ 650 m in west Timor to up to
~250 m in east, during late Miocene-early Pliocene; Figs. 3.12 and 3.13) is more likely to

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Figure 3.12. Distribution of deflection according to its cause along line 165-9 in southwest Timor. These values are based on results shown in Fig. 3.9. The last cycle (2.5 to Recent) does not have enough quality data for this analysis. Note the continuous decrease in deflection with time and change in forebulge position. Tectonic load is not at scale. The elevation of the thrust-wedge is between ~2100 and 945 m. Sedimentary loads do not create topography in Timor Sea.
Figure 3.13. Distribution of deflection according to its cause along line 118-15 in northeast Timor Sea. These values are based on results shown in Fig. 3.10. Note the increasing amount of deflection with time. Tectonic load is not at scale. The elevation of the thrust-wedge is between ~1800 and 840 m. Sedimentary loads do not create topography in Timor Sea.
be the cause of this asynchronous marine fluctuation. Transgression occurs because the subsidence rate (up to ~ 350 m/m.y.) is higher than the eustatic changes (up to 33 m/m.y.).

Regressions are not easily explained in Timor Sea, but forebulge development caused by sediments and tectonic loading is thought to be, at least in part, responsible for drops in sea level (Figs. 3.12 and 3.13). In addition, the eustatic curve (Haq et al., 1988) shows a net sea level fall since middle Miocene time (about 50 m). Therefore, periods of tectonic quiescence may have also developed marine regressions. During late Pliocene the number of marine cycles is similar throughout the basin.

Wehr (1993) shows asynchronous marine fluctuations along the Cretaceous sea belt in Wyoming as a product of differential subsidence, similar to what happens in Timor Sea. A corollary from these observations is that in foreland basins, chronostratigraphic correlations based on unconformities and maximum flooding surfaces could be erroneous, because they both can developed coevally.

Deflection produced by sedimentary loading is very significant and controls the regional width of the basin (Figs. 3.12 and 3.13). The length of this type of deflection is markedly longer than the deflections created by tectonic loading. By contrast the deflection produced by tectonic load is narrow and deep. The accommodation needed to contain late Miocene to Recent sedimentary records requires tectonic loading to have controlled only 20 to 48% of the subsidence. But in order to create today’s Trough bathymetry, it requires ~70%. The difference between these two values of subsidence is not represented in the stratigraphic record, it is represented in today’s water depth It also
implies a history of sediment starvation (i.e., accommodation is created at rates higher than sedimentation) during the entire basin evolution.

Forebulges are not distinguishable in the seismostratigraphic record. This is probably due to progressive changes in the position, distribution and magnitude of the tectonic and sedimentary loads, which cause development and overlap of subtle bulges at different times. Forebulges do not migrate, they change in time and space and sometimes they pile-up, as shown by models (Figs. 3.12 and 3.13). Results show the position of forbulges created by each type of load (tectonic and sedimentary) as well as the composite one.

In all cases, the forebulges induced by tectonics (Figs. 3.12 and 3.13) are closer to the Trench than those induced by sediments, even though the deflection is shallower. Sedimentary forebulges are invariably located in the foreland direction, closer to the continent (Australia) than tectonic bulges (Figs. 3.12 and 3.13). This is due to the distribution of the sediments over long parts of the plate that enlarges and deepens the first order deflection. The tectonic forebulge can be depressed or “deflected” along with the first-order deflection induced by sediments. The composite forebulge is located between the tectonic and the sedimentary induced bulges, or beyond the sedimentary forebulge in the foreland direction (Figs. 3.12 and 3.13). It could indicate that the tectonic forbulge is dominant in the first case, or conversely, that it is the sedimentary-induced bulge the dominant in the later case. Position of composed forebulge changes randomly in time and space when east and west Timor Sea are compared (Figs. 3.12 and 3.13). This is due to the fact that sedimentary load will change geometry according to the accommodation availability. However, when only the tectonic forebulges are considered,
it is clear that the thrust-belt moved initially as a forward breaking sequence and forbulges developed accordingly in the landward direction (Fig. 3.12, cycles between 6.5 and 4.5 Ma., and Fig. 3.13, cycles between 6.5 and 3.0 Ma.). Later, forbulges move backward towards the hinterland, indicating probably an out-of-sequence and/or backthrusting deformational events. Therefore, determinations of thrust-belt dynamics based on composed forbulges are erratic. A real correlation of forbulges and tectonic dynamic is only possible when tectonic forbulges are distinguished from those induced by sediment loading.

**Summary**

The effective elastic thickness of the Australian lithosphere in the Timor Sea is between 70 to 100 km and the overall rigidity of the plate did not change during foreland basin evolution.

Lithospheric flexure due to tectonic loading of the Banda Orogen accretionary prism is responsible for about 70% of the total subsidence observed during the development of the Timor Sea Foreland basin in the Northwest Shelf of Australia. It also caused local sea level changes within the basin. The regional isostatic compensation of sediments roughly produced the remaining 30% subsidence, and the effect of eustatic changes was almost negligible.

The ~500 km long deflection of the Australian plate in the Timor Sea is asymmetric as result of the diachronous character of the thrust progression. Thus, from ~6.5 to ~3.4 Ma the southwestern area of the Timor Sea was deflected at least 2 times as much as the northeastern area, causing local marine transgressions and regressions. Over
long periods of time, forebulges are more apt to be a broad composite feature of a foreland basin and can produce local marine regressions.

In the Timor Sea, chronostratigraphic correlations based on unconformities and flooding events might be erroneous. Local sea level fluctuations created by asymmetrical flexure are responsible for complete marine cycles of limited lateral extension. Therefore, assessment of the deflective history of the lithosphere could be crucial for the evaluation of sea level cycles in foreland basins.

References


CHAPTER FOUR. SEDIMENT AND TECTONIC INDUCED FLEXURE IN RETROARC FORELAND BASINS: EVOLUTION OF THE NORTHWESTERN-ANDES PUTUMAYO BASIN, COLOMBIA

Introduction

The regional geometry of foreland basins is the product of successive adjustments of the underlying elastic lithosphere to tectono-sedimentary loading events. These adjustments may involve changes in plate strength in time and space (Beamount 1981; Lavier and Steckler, 1997; Clark and Royden, 2000; Cardozo and Jordan, 2001). Additionally, changes in the amount, distribution and timing of loading and resultant deflection (Chapter 2) produce spatially and temporally variable, regional stratigraphic architecture. Despite this fact, temporal changes in the strength of the lithosphere and flexural response to vertical loading throughout the evolution of foreland basins remain largely ignored.

Retroarc foreland basins form on the over-ridden plates of convergent continental margins, along the interior flanks of orogenic belts, usually, magmatic arcs (Dickinson 1974; Fig. 3.1). These basins, which contain most of the global occurrence of non-marine strata, are large scale (1000’s km long), long-lived features (10 to 100 m.y.) that in cases may accumulate up to kilometers of sediments (Jordan, 1995). A typical example is the Andes fore-deep along the western margin of South America. The Putumayo sub-basin in Colombia (Fig. 4.1) has been forming, along the interior flank of the volcanic northern Andes, since latest Cretaceous time. More than 4000 m of sediment-thickness has accumulated in the basin since the first collisional tectonic pulse (Cordoba et al., 1997). Stewart and Watts (1997), using gravity data modeled the present geodynamic stage of
the lithosphere in the Putumayo-foreland. However, little is known about changes in the geometry and strength of the underlying lithosphere of the basin throughout its foreland history.

In this chapter, seismostratigraphic data of four previously recognized chronostratigraphic units in the Putumayo Basin, reduced to decompacted-thickness profiles, are forward modeled to determine any changes in lithospheric strength, and in the distribution and magnitude of supracrustal loading throughout the evolution of the basin. The amount of flexure caused by the thrust belt and sediment loads are identified independently using the method explained in Chapter 3. Results suggest that most of subsidence in the basin is caused by sediment loading. The South American elastic lithosphere in the area responds in a weak manner to supra-crustal bading as determined in previous models for the Andes (Stewart and Watts, 1997; Sarmiento, 2002).

Geologic Setting

The Putumayo Basin is part of the so-called Sub-Andean Basin System developed along the eastern margin of the Andes Mountains in South America (Sarmiento, 2001). The basin, located along the eastern flank of the Andean Cordillera in southern Colombia, extends into Ecuador and Peru with the names of Gran Cuenca de Oriente and Marañón basins, respectively (Figs. 4.2, 4.3). The Putumayo Basin has developed over Pre-Cambrian continental crust, ~35 km thick whose origin is related to the Guyana Shield in the east (Covey and Dengo; 1993Mora et al., 1998; Sarmiento, 2002; Fig. 4.3).

Three regional tectonic episodes formed the present configuration of the basin: an initial Pre-Cambrian to Cambrian passive margin episode, a subsequent back-arc rifting
period during late Paleozoic and Mesozoic times and the final, mostly Cenozoic foreland phase that is still exhibited in the basin today.

Figure 4.1. Volcanic arc along the western margin of South America (white shadow). The Andes foredeep was developed along the interior flank of the Andes. White rectangle represents area of Figures 4.2 and 4.3.
The initial, scarcely documented, Pre-Cambrian-to-late-Paleozoic rifting period created a passive-margin basin along the northwestern South American plate. According to Cordoba et al. (1997) metasedimentary rocks of this age outcrop along the Eastern Cordillera and normal faults interpreted in seismic data suggest this obscure episode (Fig. 4.4).

The second episode, a well-documented change from passive to active margin, took place from late Paleozoic to early Cretaceous time, when a magmatic arc emerged to partially form the present day Central Cordillera in Colombia (Macia and Mojica, 1983; Dengo and Covey, 1994; Cordoba et al., 1997; Jimenez, 1997; Mora et al., 1998; Sarmiento, 2002; Fig. 4.4).

This magmatic arc, that constitutes the core of the northern Andes, relates to the east-dipping subduction complex developed along the western margin of the northwestern South American plate. From latest Jurassic to late Cretaceous time magmatism ceases (Pindell and Erikson, 1993; Dengo and Covey, 1994) and a rifting event associated with the opening of the proto-Caribbean, as North America and South America begin to separate (Dengo and Covey, 1993), extends into southern Colombia. The effect of these events in the Putumayo basin are relatively minor. Shallow marine deposits of transitional to inner platform facies comprise most of the Cretaceous stratigraphic section in the basin (Mora et al., 1998).

The third tectonic episode marks the beginning of the foreland period in the Putumayo area during latest Cretaceous-early Paleogene time. The oceanic crust of today’s Western Cordillera (Figs. 4.3 and 4.4) is accreted to the Central Cordillera (Henderson, 1979; Aspend et al., 1987), producing a thick-skin styled thrust belt and
tectonic inversion of Paleozoic normal faults along the eastern margin of the basin (Sarmiento, 2002). The mostly continental Rumiyaco Formation is deposited during this initial collisional period (Fig. 4.4).

Figure 4.2 Location of Putumayo Basin. Note the crystalline Precambrian Guyana Shield and Vaupes Arch (After Jimenez, 1997).
Figure 4.3. Modern tectonic framework of Colombia. Large-dark arrows indicate relative movement, with adjacent rates, between each plate. Numbers represent GPS stations as follow: (1) Villavicencio, (2) Bogotá, (3) Malpelo, (4) San Andres Island. (After Jimenez, 1997).
Figure 4.4. Tectonic evolution model of the Putumayo Basin from Cambrian time (not at scale) (Modified from Cordoba et al., 1997)
The foreland evolution of the Putumayo Basin during Tertiary time is characterized by three other significant tectonic pulses related to tectono-sedimentary units (Cordoba et al., 1997): First, a late Paleocene-middle Eocene compressive pulse originated in response to an increasing in the convergence rate between the Nazca and South American plates (Daly, 1989). The continental Pepino Formation was deposited mostly in alluvial fans as product of continuous uplift of the Central Cordillera (Mora et al., 1998; Fig. 4.5). This unit is topped by a regional angular unconformity that, on the west side of the basin, cuts thrusts and back-thrusts developed during this period. Second, the collision between the Andes and the Chocó Terrene produces a new tectonic pulse that prolongs the exhumation of the Central Cordillera (Sarmiento, 2001). This took place during Middle Eocene to Middle Miocene and is responsible for the continental Oligocene Orteguaza Formation as well as the Miocene Orito and Belen Formations. The last tectonic pulse of the foreland stage of the basin, produced by the collision of Panamá with Colombia during late Miocene to Recent times, generates the greatest shortening event and present uplift of the Central Cordillera as well as the development of the double vergent thrust belt than constitutes the Eastern Cordillera, north of the Putumayo (Figs. 4.3 and 4.4). The Putumayo basin area remains as an active foredeep depocenter for sediments from the Central Cordillera. Fluvial deposits of this period have been informally called Ospina and Caiman Formations.

Data

Seismic data is the primary source of information for this study. Published data and abundant well information are also used to constrain input parameters during modeling (Fig. 4.6). Over 700 km of 2D proprietary seismic data from Ecopetrol
(Colombian State Oil Company), 3 seconds in length, are interpreted. Composite lines 3 and 5 (assemble by adding more than one single seismic survey) oriented in an east-west direction are used for flexural analysis (Fig. 4.6). Thickness of units affected by the folding and faulting is not used for forward modeling.

These lines represent the east-west regional geometry of the Putumayo Basin. Over 30 wells (proprietary reports of Ecopetrol, Appendix 5) containing abundant paleontological data are used for interpretation of isochore maps. Maps of Rumiyaco Formation (Fig. 4.7), Pepino Formation (Fig. 4.8), and Orteguaza-Belen-Orito Formations (Fig. 4.9) show the lateral extent and thickness changes of the tectonostratigraphic units used to constrain modeling. Invariably all units have a wedge-shaped geometry (Figs. 4.7, 4.8 and 4.9). These geometries are used during modeling to constrain the shape of the sedimentary loads. Well information was used by Geotec (1992) and Cordoba et al. (1997) to constraint the unconformities that divide the four tectonostratigraphic units. Abundant fossil-records from well cores have been used for age determination and time-gap estimates.

**Data Reduction**

Seismic data and isochore maps show the regional wedge-like geometry of the sedimentary record during the entire foreland period of the Putumayo basin. Lithostratigraphic thickness calculation is based on time-depth curves estimated from over 30 wells (Geotec, 1998; Fig. 4.10) and controlled with stratigraphic well information where possible. Decompaction of the sediment packages to initial conditions for each cycle was carried out according to Sclater and Christie (1980) and Watts (2001)
(Appendix 3). According to Cordoba et al. (1997) these conditions include an original porosity of 48%.

Figure 4.5. Generalized stratigraphic column of the Putumayo Basin (Modified from Geotec, 1992). Note the unconformities that are interpreted along the seismic data. Ss/Sh is sandstone/shale ratio.
The porosity-exponential constant is 0.27/km (Cordoba et al., 1997), the grain density was taken as 2700 kg/m$^3$ and the density of infill sediments used was 1807 kg/m$^3$ (Appendix 3). Thickness reduction by compaction is estimated to lie in the 15 - 35% range. Structural shortening across the Andes thrust-belt in the Putumayo Basin has been estimated not larger than 15% since Paleocene time (Jimenez, 1997). Thickness values from isochore maps are assumed to be the maximum for each unit and used for decompaction. The error in estimated thickness introduced by this assumption, and the amount of shortening is 10 to 13%.

Figure 4.6. Seismic and well information in the Putumayo Basin. Chronostraigraphic data in well-penetration chart (Appendix 5)
Figure 4.7. Rumiyaco isochore map (Modified from Cordoba et al., 1997).

Figure 4.8. Pepino isochore map (Modified from Cordoba et al., 1997).
Figure 4.9. Orteguaza-Orito-Belen isochore map (Modified from Cordoba et al., 1997).

Figure 4.10. Depth-Time curve based on well information (After Geotec, 1992).
Stratigraphy of Tectono-Sedimentary Units

Models in non-marine foreland basin sequence stratigraphy focus on either the control exerted by eustasy or tectonics in the fluvial sedimentation and regional stratigraphic architecture. Classical sequence-stratigraphy concepts have been apply to the middle to upstream portions of fluvial systems without much success, because the role of sea-level in creating or destroying accommodation in these deposits is not clear (Posamentier and Vail, 1988; Shanley and McCabe, 1994; Emery and Myers, 1997). It is thought that fluvial systems evolve in response to a set of allocyclic and autocyclic processes such as river-profile slope, sediment load, bed-rock material, climate and tectonic regime (Schumm and Ethridge, 1991; Emery and Myers, 1997). These controls may change abruptly or gradually, over long periods of time (>10⁶ m.y.) (Schumm et al., 2000). Therefore, the response of the fluvial-system stratigraphy is erratic at any geological scale and varies according to where along the river profile the processes take place (Schumm and Ethridge, 1991). Schumm (1977) divides the fluvial system into three different geographical and geological areas (Fig. 4.11). First, in the upstream area (hinterland and source area for the river), tectonics plays a significant role, not only in source and size of sediments, but also in the regional architecture of the deposits, changes in the shape of graded profile and in accommodation regime. The second geographical area is the mid-stream or the segment of drainage where sediment is transported. The third is the downstream or the lower reaches of the river, near coast lines and deltas. Autocyclic processes and eustasy, respectively, are thought to be the most significant factors controlling fluvial dynamics and sedimentation.
Shanley and McCabe (1994) discuss the relative importance of downstream base level controls versus upstream tectonic controls in the development of fluvial architecture (Fig. 4.12). In general, it has been accepted that the importance of change in base level (sea level) diminishes upstream (Shanley and McCabe, 1994; Schumm et al., 2001). For example, sea level rise and fall change fluvial architecture 220 km and 90 km upstream along the Mississippi and Colorado rivers, respectively (Shanley and McCabe, 1994). They also recognize sequence boundaries in fluvial systems as incised valleys eroded during sea level fall and well-developed interfluve paleosoils. Lowstand to transgressive systems tracks are represented by braided to meandering styles respectively that reflect changes in accommodation (Miall, 1997).

4.11. Fluvial-systems model showing three geomorphic areas with different allogenic controls (After Schumm et al., 2001)
The concept of graded river developed by Mackin (1948) establishes that the regional slope of a stream tends to adjust to a profile that corresponds to the channel flow-velocity required to transport the sediment load supplied to the basin; i.e., there is no net erosion or deposition along the profile. This equilibrium surface reflects the influence of numerous local base-levels, including lakes, trunk-stream drainage, position of nick-points, position of ground-water table and sea level (Wescott, 1993; Miall, 1997; Emery and Myers, 1997). For this reason the concept of graded stream profile has been applied to fluvial systems to define the equilibrium surface that controls accommodation (Mackin, 1948; Sloss, 1962; Emery and Mayers, 1997). Most papers dealing with sequences in fluvial systems show how the stream profile is disrupted by base-level changes along the downstream portion of the system, where sea level controls.
stratigraphic architecture in a similar fashion to classical sequence models (Posamentier and Vail, 1988; Shanley and McCabe, 1994; Miall, 1997). However, few papers deal with graded profiles disrupted along the upstream portion of the system, as for example, by tectonic uplift, which is critical in non-marine retroarc foreland basins, formed mostly due to flexure of the plate under supracrustal loading.

Posamentier and Vail (1988) developed a model for alluviation during the last part of high stand system tract. The equilibrium profile of the river will vary according to eustatic changes. The stream system will aggrade if its equilibrium profile shifts basinward or upward creating new horizontal or vertical accommodation, respectively. Conversely, it will incise (erode) if its equilibrium profile shifts downward. However, they argue that the portions of the river affected by these changes are the lower parts of the profile. Additionally, they assume constant or non-subsidence along the continental portion of the system, where fluvial systems develop.

Tectonic uplift in non-marine foreland basins affects lithofacies distribution in fluvial sedimentation and changes in erosion/deposition rates (Schumm et al., 2000). Extrabasinal uplift, considered the primary source of sediments for fluvial systems, increases slope and consequently the river competence and sediment loads. Source-area uplift thus tends to result in gravel progradation in to the basin (Ibid.). Discussion exists about the synchronicity between uplift and conglomeratic deposits (Flemings and Jordan, 1989). Early studies establish a direct link between conglomeratic strata and source area uplift (Rust and Koster, 1984). However, recent observations show that because rates of tectonic uplift on average are eight times higher than average denudation rates (Blair and McPherson, 1994), a lag time exists between uplift and gravel deposition whereby
progradation occurs well after a major tectonic pulse. At times when significant relief exists between the uplifted belt and the subsiding basin (Fig. 4.13), the tilted plate tends to develop a trunk-axial drainage parallel to the linear load (e.g. thrust belts). During these periods, the primary sediment source is from cratonic areas (Lawrence and Williams, 1987; Fig. 4.13). Once the sediment supply from uplifted belts is re-established, it will occupy the available accommodation and the downstream slope is reinstated. On the basis of the amount of sediment supply, Jordan (1995) divided the evolution of non-marine foreland basins in two stages: the first, underfilled stage, when the axis of the depocenter is parallel to the thrust-belt (Fig. 4.13). The second, overfilled stage, when the accommodation is totally filled and the sediment, coming from the thrust belt, bypasses the forebulge and transport continues downstream.

Figure 4.13. Concept of underfilled (A) and filled (B) non-marine foreland basin (After Jordan, 1995).
Subsidence in non-marine foreland basins generates space for fluvial deposition through lowering the basin floor (Schumm et al., 2000). Experimental models developed by Allen (1978) show that slow subsidence promotes channel belt occurrence and deposition of multi-storey and multi-lateral sandstone sheets with high sandstone/mudstone ratios (Fig. 4.14).

Rapid subsidence, by contrast, permits significant deposition of overbank “fines” before the re-initiation of channel belts at a new location, resulting in a low sandstone/mudstone ratios (Ibid.) (Fig. 4.14). However, sandstone/mudstone ratio changes may also indicate facies changes from braided (high-ratio) to meandering (low-ratio) as well as changes in sediment supply. Sea-level fluctuations, in downstream
portions of fluvial systems, can generate accommodation similar to that created by updip subsidence (Schumm et al., 2000).

At regional scale, Jordan and Flemings (1991) and Flemings and Jordan (1989) developed sequence stratigraphy models for foreland basins characterized by sedimentary wedges during tectonic pulses and lens-like packages during tectonic quiescence (Fig. 4.15). Similarly, Catuneanu et al. (1997) generate comparable geometry by invoking dynamic forces and “reciprocal stratigraphy”: wedge-like geometries during tectonic activity and lens-like geometry during dynamic deflection. They also argue that due to the dynamic subsidence induced by sediment weight, forebulges subside along with the entire margin and therefore are hard to identify. Puigdefabregas et al. (1986) show models of regional stratigraphy characterized by clinoforms with sediment sources toward the hinterland.

Figure 4.15. Sequence stratigraphic model for non-marine foreland basin according to Jordan and Flemings (1991), and Jordan (1995). \( Y_1 \) and \( X_1 \) represent periods of tectonic activity. \( Y_2 \) and \( X_2 \) represent periods of tectonic quiescence (not at scale).
DeCelles and Giles (1996) define a foreland system as an elongated region of potential sediment accommodation that forms on continental crust between a contractional orogenic belt and the adjacent craton. It consists of four depozones (Fig.4.16): wedge-top, foredeep, forebulge and back-bulge zones. They suggest that sequence stratigraphic models must include a doubly-tapered prisms rather than the typical wedge-shape geometry. However, their model assumes that the “primary” forebulge is created by tectonic loading and based on its position, they recognize the other three components of the system.

Figure 4. 16 Foreland basins system according to DeCelles and Giles (1996). Note the doubly-tapered prism and the four depozones.
An attempt to identify the geometry of sequences of the tectono-stratigraphic units at regional-scale as well as at seismic-scale is carried out on lines 3 and 5. In agreement with established seismo-stratigraphic models (e.g., Jordan and Flemings, 1989), the regional geometry, i.e., wedge-like or lense-like bodies, is highlighted. In addition, seismic lines are interpreted within a standard seismic stratigraphic approach to reflector geometry, at the scale of the seismic data. However, it is necessary to keep in mind that the foredeep deposits of the Putumayo basin represent mostly continental records of fluvial origin (Cordoba et al., 1997). Thus, the equilibrium profile concept is used as the slope to which the river-profile tends to be adjusted. It is considered that large-scale fluvial sedimentary sequences are developed during this process (Emery and Mayers, 1997) and therefore it is as critical as sea level acting as the effective upper boundary of accommodation (Emery and Mayers, 1997; Schumm et al., 2000). Additionally, to determine whether any relationship exists between sea level and foreland sedimentary cycles and sequence development in Putumayo, a comparison between the numbers of third-order sequences established by Haq et al. (1988) and the sedimentary cycles found along the data (defined by onlap shifting) is carried out.

The four previously identified tectono-sedimentary units (Rumiyaco Formation, Pepino Formation, Orteguaza-Belen-Orito Formations, and Ospina-Caiman Formations) separated by regional unconformities are interpreted (Figs. 4.17 and 4.18). Regionally, the four tectonostratigraphic units exhibit the typical wedge-like geometry. The well-documentated unconformities do not show angular relationships such as truncations or abrupt (angular) reflector terminations against the unconformity in the data-window. Interpretation is focused along undisturbed areas not affected by thrusting at any time,
therefore conserved in almost pristine stage. The vertical scale is defined by seismic resolution (velocity field). Published information (Geotec, 1992; Cordoba et al., 1997; Fig. 4.5) is used to determine the sandstone/shale ratio.

A brief lithologic description of each unit, as well as depositional environments, time represented by the succession, the number of marine sequences that appear in the eustatic chart (Haq et al., 1988) during the elapsed time of each section and the seismic facies exhibited, are shown below as a comparison tool to elaborate possible evolution of sequence development in the Putumayo basin.

**Rumiyaco Formation (Maastrichtian-Late Paleocene)**

This unit contains mostly gray shale and some interbedded sandstone layers toward the base and reddish siltstones and mudstones at the top. This formation represents the transition from shallow marine environments to continental conditions (Fig. 4.5; Cordoba et al., 1997). The isochore map (Fig. 4.7) shows the thickness of this unit reaching ~ 1200 meters at Bagre-1 and Gavilan 1A wells (Figs. 4.6 and 4.7; Cordoba et al., 1997). The maximum decompacted thickness according to seismic data is 1264 m. The maximum sedimentation rate is ~46 to 63 m/m.y. The sandstone/shale ratio is 0.17. The age of this unit is based on palinologic studies by Mora et al. (1998) that report the association *Buttinia andreevi* (Maastrichtian), *ulmoideipites kreempii* (Maastrechtian to Eocene), *Bambacacidities bellus* (Eocene) and *Polipodiaceaisporites sp.* (Eocene). The underlying discordant contact has been interpreted in several wells along the basin (Appendix 5).
Figure 4.17. Seismic line 3 (see Fig. 6 for location). Uninterpreted (above) and interpreted section. Note the deformational front at the west-end of the basin (thrust faults).
The eustatic curve (Haq et al., 1988) shows that between the base of the Maastrichtian (~ 74 Ma) and the top of the Thanetian (~54 Ma) there are 8 third-order sequences, with a net sea-level fall of ~60 m during the entire elapsed time (sea-level fall rate ~3 m/m.y.). According to Vail et al. (1977) basinwardshifts in onlap determine the number of sequences.

Seismic data exhibits reflections of high lateral continuity (> 60 km) that decreases upward to moderate lateral continuity (20 to 60 km). The amplitude of this reflector is moderate to low (decreasing upward). Acoustically, this unit is semitransparent with parallel to oblique reflectors geometry.

The base of this tectonostratigraphic unit is marked by four onlap terminations against the basal unconformity in the foreland direction (Figs. 4.19 and 4.20). Each individual unit, topped by an onlapping event, shows a wedge like geometry, which indicates differential subsidence rates of the depositional surface. Line 3 (north) shows some basinward onlap shift with one associated offlap event (progressive basinward migration of the stratigraphic units). The interpretation is that it reflects basin deepening closer to the tectonic load. Offlapping can indicate rapid subsidence and sediment source locate in the foreland.

**Pepino Formation (Late Eocene-Early Oligocene)**

Two conglomeratic units, separated by a succession of gypsum and limolitic siltstones, constitute this unit (Fig. 4.5, Cordoba et al., 1997). Alluvial fans and braided streams are the depositional environments that have been interpreted for the conglomeritic units (Portilla et al., 1993). For the middle silty unit, alluvial plains and outer alluvial fans have been interpreted (Portilla et al., 1993). This formation records the
Central Cordillera uplift, identified as sediment source area. Two tectonic pulses that correspond with the conglomeratic units characterize this event. (Cordoba et al., 1997).

The isochore map (Fig. 4.8) shows the thickness increasing from ~ 100 m (Azul Curbinata-1 well) in the east to ~550 m toward the west (Azul Grande-1 well) (Fig. 4.8), showing a clear west-dipping paleo-slope direction during depositional time. From seismic and well data the maximum estimated decompacted thickness is ~680 and sedimentation rate for this unit is ~68 m/m.y. The sandstone/shale ratio is 0.8. Mora et al. (1988) report *Verrucatosporites* cf. *usmensis* and *Cicatricosisporites* cf. as evidence of the late Eocene age of this Formation. The underlying discordant contact has been interpreted in several wells along the basin (Appendix 5).

The eustatic curve (Haq et al., 1988) shows that between the base of the late Eocene (~ 40 Ma) and the top of the early Oligocene (~ 30 Ma) there are 6 third-order sequences, with a net sea-level fall of ~ 30 m, during the entire elapsed time (sea-level fall rate ~ 3 m/m.y).

Seismically this unit consists of moderate to high reflection lateral continuity (increasing upward). The amplitude of this unit increases from low to high from base to top. Acoustically, this unit is semi-transparent toward the base but becomes a succession of parallel-conformable events to a discontinuous series of oblique/sigmoid reflectors with one clear toplap termination against the upper-boundary unconformity. The underlying unconformity is a large-scale feature that extends over 100 km. Basal reflections onlap the underlying basal unconformity (Fig. 4.20). Differences between the lower and upper part of the tectonic unit might reflect an increase in subsidence rate close
to the hinterland. Downlap terminations toward the hinterland might indicate that sediment source is the craton toward the foreland.

**Orteguaza- Belen-Orito Formations (Late Oligocene-Late Miocene)**

Stratigraphically, these three units are difficult to differentiate from each other using well data (Cordoba et al., 1997). The lower unit (Orteguaza Fm.) contains shale with interbedded fine-grained sandstones layers and some coal beds typical of paralic environments. The upper two units contain a monotonous sequence of red-gray shale with some interbedded layers of sandstones and conglomerates and occasional gypsum and coal. Depositional environment has been interpreted as mostly fluvial with some transitional layers (Portilla et al., 1993). Figure 4.9 shows the extend of this unit (over 300 km) and its thickness changes which increases steadily from ~800 m from the east to ~ 1300 m toward the west (Oil fields Burdine, Nancy ad Maxine; Cordoba et al., 1997). Maximum estimated decompacted thickness from seismic and well data is ~1900 m and maximum sedimentation rate is estimated in 145 m/m.y. (Appendix 3). The sandstone/shale ratio is 0.2 (Fig. 4.5). Rubio (1997) report a Oligocene to Miocene age. However, Germeraad et al. (1967) restrict the range from late Oligocene to early Miocene, based on the palinological zonation *Cicatricosisporites dorogensis* and *Verrutricolporites rotundiporus*. Zone *C. dorogensis*, and parts of Zone *V. rotundiporus*.

The eustatic curve (Haq et al., 1988) shows that between the base of the late Oligocene (~ 27 Ma) and the top of the Middle Miocene (~ 13 Ma) there are 11 third-order sequences, with a net sea-level fall of ~50 m during the entire elapsed time (sea-level fall rate ~2.5 m/m.y.).
Figure 4.18. Seismic line 5 (see Fig. 6 for location). Uninterpreted (above) and interpreted section. Note the deformational front at the west end of the basin.
Figure 4.19. Seismic expression of foreland basin along Line 3. Thicker lines are dated hiatuses. Note the onlap basinward shifting that indicates basin deepening. Downlapping toward the basin indicates sediment source from the craton.
Seismically this unit exhibits reflections of high lateral continuity that decreases upward to moderate lateral continuity. The amplitude of this unit is high and the frequency is the highest along the data-survey (14 or more reflections in ~ 450 ms; Figs. 4.18 and 4.19). Acoustically this unit shows a succession of parallel reflections with some low obliquity toward the top. The underlying unconformity is a large-scale feature that extends over 100 km. Along line 5 (Fig. 4.20), two consecutive downlap reflectors mark the beginning of the cycle, however in the north the first reflection termination is an

Figure 4.20. Seismic expression of foreland basin along Line 5. Note the onlap basinward shifting. This indicates basin deepening. The Orteguaza-Belen-Orito tectonostratigraphic sequence shows downlap in the basin direction, this can indicate sediment coming from the craton (east), and probably an underfilled stage of the basin. For location see Fig. 4.18.
isolated downlap event that appears overlying two parallel toplap reflectors that end beyond the data-window. Lines 5 (south) exhibits two onlap shifting-events in the hinterland direction, separated by a downlap reflector, near to the middle of the sequence (Figs. 4.19 and 4.20). Toward the top, both sections show a zone of short downlapping reflectors that at times appear chaotic, punctuated by the last onlap shift of the cycles. The remaining events, that appear plane-parallel in the survey, end outside of the data-window.

**Ospino-Caiman Formations (Middle Miocene – Recent)**

Little information exists about these units. It is thought that they record the last tectonic pulse of the Andean Orogeny containing numerous successions of conglomerates and coarse sandstones bodies typical of alluvial fans and braided systems (Cordoba et al., 1997). Maximum decompacted thickness is ~ 1600 m and sedimentation rate is ~142 m/my and sandstone/shale ration is 0.3. Rubio (1997) reports a late Miocene to Recent age, whereas Germeraad et al. (1967) dated these units as middle Miocene to Recent, based on palinologic Zone *Verrutricolporities rotundiporis (Psiladiporites minimus)*, and *Verruscatosporites usmensis*.

The eustatic curve (Haq et al., 1988) shows that between the base of the early Pliocene (~ 11.2 Ma) and Recent time there are 7 third-order sequences, with a net sea-level fall of ~30 m during the entire elapsed time (sea-level fall rate ~ 6 m/m.y.).

Seismically only the lower part of this unit shows identifiable seismic facies: moderate to high lateral reflector continuity, with moderate to high amplitude. Acoustically this unit is a semitransparent sequence of parallel reflectors that onlap the top of the previous cycle (Figs. 4.19 and 4.20). However, no other reflector-geometry is
discernable due the poor quality of the seismic data. The underlying unconformity is a large-scale feature that extends over 100 km.

**Flexural Modeling**

Implementation of the model is similar to that used in chapter three (Fig. 3.5). However in the Putumayo area eustasy is assumed not to have any direct influence in the resultant accommodation, as the continental nature of the sedimentary record rules out any marine presence. Modern deposits along the Putumayo piedmont completely fill the basin and paleo-sources analyses suggest sediment transport has been always perpendicular to the axis of the basin (Cordoba et al., 1997). It is assumed that the basin behaves as an overfilled depocenter, i.e., the entire accommodation is full of sediments during the entire foreland period.

The four tectonostratigraphic events that have been matched with lithostratigraphic units, namely Rumiyaco, Pepino, Orteguaza-Orito-Belen, and Ospino-Caiman Fms. (Fig. 4.5) are used for flexural analysis. The geodynamic constants used during modeling are in Table 4.1.

The critical taper angle was estimated from published seismic and structural sections across the Andes thrust belt in Putumayo, and Napo basin in Ecuador (Table 4.2). Density of the thrust-wedge is taken from published data (Cordoba et al., 1997). The dominant thick-skin tectonic style (Jimenez, 1997) implies crystalline rocks are involved in some of the thrust sheets. 1000 kg/m$^3$ change in density implies a change in deflection of ~100 m, using a plate with 100 km of effective elastic thickness.
Table 4.1. Geodynamic constants used for modeling. Values taken from Turcotte and Schubert (1982), and Cordoba et al. (1997). These values are kept constant during modeling.

<table>
<thead>
<tr>
<th>Constant</th>
<th>Symbol</th>
<th>Value</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gravity acceleration</td>
<td>g</td>
<td>9.8</td>
<td>m/ s²</td>
</tr>
<tr>
<td>Water density</td>
<td>ρₚw</td>
<td>1035</td>
<td>kg/m³</td>
</tr>
<tr>
<td>sediment density</td>
<td>ρₛ</td>
<td>1800-1900</td>
<td>kg/m³</td>
</tr>
<tr>
<td>Density of tectonic wedge</td>
<td>ρₛw</td>
<td>2500</td>
<td>kg/m³</td>
</tr>
<tr>
<td>Density mantle</td>
<td>ρₚm</td>
<td>3300</td>
<td>kg/m³</td>
</tr>
<tr>
<td>Young Modulus</td>
<td>E</td>
<td>5x10¹¹</td>
<td>Pa (kg/m·s²)</td>
</tr>
<tr>
<td>Poisson’s ratio</td>
<td>ν</td>
<td>0.25</td>
<td></td>
</tr>
</tbody>
</table>

For modeling the tectonic wedge (Appendix 2), a critical taper angle of 7 degrees was considered. It was the approximated average from sections based on seismic data. Interpretative sections were not considered (Table 4.2).

Table 4.2. Thrust-wedge taper-angle values.

<table>
<thead>
<tr>
<th>Author</th>
<th>Taper Angle</th>
<th>Year</th>
</tr>
</thead>
<tbody>
<tr>
<td>Portilla</td>
<td>6</td>
<td>1991</td>
</tr>
<tr>
<td>Dengo (Llanos)</td>
<td>8</td>
<td>1993</td>
</tr>
<tr>
<td>Balkwill et al.</td>
<td>6</td>
<td>1995</td>
</tr>
<tr>
<td>Jimenez</td>
<td>10</td>
<td>1997</td>
</tr>
<tr>
<td>Cordoba et al.</td>
<td>6</td>
<td>1997</td>
</tr>
<tr>
<td>Geotec</td>
<td>7</td>
<td>1992</td>
</tr>
</tbody>
</table>
Model Results

The effective elastic thickness of the continental lithosphere in the Putumayo Basin is between 20 and 40 km. This value is higher than previous calculations for areas near the thrust belt but lower than previous calculations for areas closer to the craton (Stewart and Watts, 2000). The strength of the plate is similar to that of the Bolivian elastic lithosphere along the Andes thrust belt (Stewart and Watts, 2000). The plate rigidity did not change during the evolution of foreland basin.

At least 400 km of Colombian lithosphere were flexed (as part of the first order deflection) during Tertiary time in the Putumayo Basin (Figs. 4.21 and 4.22). The amount of flexure was similar across the strike of the basin for each tectonic pulse. Early during the initiation of the foreland basin, during Rumiyaco Fm. time (latest Cretaceous to early Paleocene), the maximum flexure is about 1264 m ± 150 m (Figs. 4.21 and 4.22), the rate of deflection is estimated up to 63 m/my ± 5m; 11m/m.y. due to tectonic loading and the remaining 52 m/m.y. due to sediment loading. During Eocene time (Pepino Fm.) the maximum flexure is estimated at ~680 m ± 100 m and the rate of deflection is up to 68 m/my ± 10 m, 16 m/m.y. due to tectonic loading and 52 m/m.y. due to sediment loading. At Oligocene-Miocene time (Orteguaza, Orito, Belen Fms.) the maximum flexure was 1650 m ± 230 m and the deflection rate was estimated in ~135 m/m.y., 30 m/m.y. due to tectonic loading and 105 m/m.y. due to sediment loading. Finally during Miocene to Recent time (Ospina Fm.) the maximum flexure was 1620 m/y ± 32 m and the deflection rate is 142 m/m.y. ± 3 m, 30 m/m.y. due to tectonic loading and 112 m/m.y. due to sediment loading. Clearly, the Andean Orogeny is responsible for the
increment in the rate of subsidence. The northern region of the basin shows greater maximum deflection than in the south. This may reflect some influence of the tectonic loading of the Eastern Cordillera that is developed during this tectonic pulse (Sarmiento, 2001; Fig. 4.3). The highest curvature ($6.5 \times 10^{-8}$ m) was reached in the northern region during Miocene time. The results also show that tectonic loading is maximum during Oligocene time. During that period the thrust-wedge responsible for the loading was 35 km long and the maximum height was ~3300 m. The minimum thrust-wedge length is 12 km and ~1500 m high during Eocene times.

Figures 4.23 and 4.24 show the total deflection and the independent contribution of each component (sediments and tectonic load) of the four tectonostratigraphic cycles. On average 22% of the flexure is caused by tectonic loading while sediment loading causes the remaining 78%.

The lowest amount of tectonic loading (17%) takes place in the northern portion of the basin during Miocene time. The highest amount of flexure due to tectonic loading is attained during Oligocene/Miocene time (26%). Deflection rates due solely to tectonic loading reach their highest level during the Miocene (~30 m/m.y.) and the lowest during the Paleocene (11 m/m.y.).

The length of the sediment load is between 190 km, during the Eocene Pepino Fm. and 410 km during Oligocene-middle Miocene Orteguaza-Belen-Orito Fms. There is not relationship between the lateral extent of the sediment load and the total flexure of the basin. The lateral extent of the load reaches its minimum during the Eocene and its maximum during Oligocene. However the amount of tectonic related subsidence is at its highest level throughout while the rate of flexure induced by sediments fluctuates without
Figure 4.21. Flexural model along seismic line 3 (see Fig. 4.6 for location). Geometry of the loads in Fig. 4.23.
Figure 4.22. Flexural model along seismic line 5. Geometry of the load in Fig. 4.23.
reaching its maximum or minimum level during this time. The amount of sediment loading \(10^{11} \text{ N/m}\) is of the same order of magnitude as that produced by tectonic loading.

The width of the basin produced by the first order sediment-related flexure, is 60 to 120 km larger than that produced by tectonic loading during a single cycle, i.e., 23 to 35% longer. The position of the forebulge produced by tectonic loading is closer to the thrust belt, higher (102 m) and narrower (~150 km) than the forebulges produce by sediment load (up to 30m high and 180 km wide).

However, no conspicuous forebulges are discernible along the composed deflection. The tectonic-related forebulge is located along the slope the first order deflection induced by sedimentary-flexure, whose forebulge is very subtle (Figs. 4.23 and 4.24). Forebulges are wide structures that overlap and produce a broad and less noticeable resultant deflection beyond the first inflection-point of the tectonic-flexure.

**Discussion**

Results are limited by the parameterization of the vertical loads. Even though there is a reasonable good control in sediments geometry, such a control does not exist in the tectonic load. Thick-skin tectonic implies for example, that crystalline rocks are emplaced within the thrust sheets, however there are not available reports about the nature of these bodies and specific densities. Changes in 100 kg/m\(^3\) in density will increase up to 8% the resultant compensation. Present day values of porosity and compaction history are considered as the result of the mechanical expulsion of water, without taking in to account diagenesis, which may reduce or increase porosity. The geometry of sedimentary loads is constrained by wells (isochore maps) with not lateral
Figure 4.23. Deflection-cause distribution along seismic line 3 (see Fig. 4.6 for location). Note the changing location of the forebulge and the difference in wavelength of the deflection due to tectonics versus sediment loading. Tectonic load is not at vertical scale. The elevation of the thrust-wedge is between ~3600 and 1260 m. Sedimentary load represents thickness.
Figure 4.24. Deflection-cause distribution along seismic line 5 (see Fig. 4.6 for location). Note the changing location of the forebulge top and the difference in wavelength of the deflection due to tectonics versus sediment loading. Tectonic load is not at vertical scale. The elevation of the thrust-wedge is between ~3100 and 1400 m. Sedimentary load represents thickness.
continuity, or by seismic data (limited regional extend). The Orteguaza-Belen-Orito unit shows little thickness changes along the seismic profile, however its isochore map clearly exhibits its wedge-like geometry. In this case, we used mostly the isochore map to constraint the load-geometry, because it shows the regional distribution of the sedimentary unit. The upper boundary of accommodation is the equilibrium profile, which quantification is limited to the sediment pile-up above the point of zero deflection (Figs. 4.23 and 4.24) is not constrained by any other data-set. Unfortunately, no palinspastic or paleotopographic maps are available. Certainty exists, however, about the continental nature of the entire foreland sedimentary succession. The sediment excess shown in Figures 4.23 and 4.24 represent paleotopography.

Regional isostatic compensation of the sedimentary cover induces the creation of up to 78% of the total accommodation in the Putumayo retroarc basin. Tectonic loading creates the remaining 20% and eustasy appears to be negligible in non-marine foreland basin. This agrees with the amount of accommodation induced by sediment in passive margin basins (Reynolds et al., 1991). However this contradicts arguments in favor of the influence of eustasy in non-marine foreland basins as proposed by Weimer (1986), since Haq et al. (1988) eustatic cycles have no evident relation with sedimentation in the non-marine Putumayo foreland basin, which is the up-stream portion of the fluvial system. Subsidence rate in the foreland is up to two orders of magnitude higher due to sediment loading and one order of magnitude higher due to tectonic loading than the eustatic curves. Orteguaza-Belen-Orito unit is peculiar in the seismic record (little thickness changes) and produces the widest deflection in our models. This period coincides with the maximum exhumation of the Central Cordillera (Andes) in Colombia (Cordoba et al.,
1997). This tectonostratigraphic period apparently is controlled mostly by sediment input that enlarged the basin (to ~500 km), therefore, longer wavelength must be expected and less conspicuous sediment thickness changes occur.

Total foreland-sediment pile-up above point of zero deflection reaches between $1100 \pm 150$ m (Line 3, Fig. 4.23) and $1350 \pm 100$ m (Line 5, Fig. 4.24). Tectonic deflection during the entire foreland period reaches ~1200 m (line 3, Fig. 4.23) and ~1000 m (line 5, Fig. 4.24). The difference between sediment pile-up and tectonic deflection is between 150 and 250 m. This difference is of the same order of the present day topography, which is about 500 to 250 m at the toe of the thrust-belt. The continental character of the basin during its foreland period indicates that it has been an area above sea level during its evolution due to sediment accumulation. Putumayo was a shallow marine basin before the onset of the foredeep depocenter (Cordoba et al., 1997). In addition, it is possible to infer that the base-level (or equilibrium profile) was above sea-level. It is the only way to preserve sediment records in Putumayo.

Both, the unconformity underlying Rumiyaco represents a hiatus of ~19 m.y. (Cordoba et al., 1997) and the unconformity underlying the Pepino unit represents a hiatus of ~13 m.y. (Ibid.), take place during times that, according to the Haq et al. (1988) chart, correspond to a eustatic maximum (supercycles UZA-4 and TA-4 respectively). According to Posamentier and Vail (1988), Shanley and McCabe (1994) and Miall (1997), eustasy would increase the accommodation on a worldwide basis, which is not the case in the Putumayo Basin. However the Pepino-Orteguaza paraconformity, that represents ~ 6 m.y (Geotec, 1992; Cordoba et al., 1997) and the Orito-Ospina paraconformity occurs during a sea level fall and coincides with a reduction of
accommodation. Hiatuses, such as those in the Putumayo basin with no evidence of structural deformation (angular unconformities) (Germeraad et al., 1967; Cordoba et al., 1997, Appendix 5), or erosion (disconformities), probably indicate times when fluvial-system profiles are near equilibrium.

Similar to marine records, non-marine foreland basin seismostratigraphy exhibits basinward onlap shift, thought it is not-related to sea level changes. These shifts probably reflect changes in the subsidence rate near to the thrust belt, due to tectonic loading. If there is no change in the rate of sediment supply into the basin, an under-filled basin can develop. Foreland onlap propagation reinitiates as the sediment supply fills the previously created accommodation. Previous models invoke viscous relaxation as the origin for deepening and onlap shifting and off-lapping, however elastic models generate similar stratigraphic architecture (e.g., Quinland and Beaumont, 1984). Sigmoid or downlap toward-the-hinterland reflector geometry can indicate the periods of underfilled basins with sediment source in the foreland (craton). It is difficult, however, to extend the nomenclature of passive-margin sequence-stratigraphy to non-marine foreland basins, since the former is almost exclusively related to sea level fluctuations and resultant construction or destruction of accommodation.

Posamentier and Vail (1988) argue that alluviation occurs during highstand systems tract as the base level shifts basinward and sub-aerial accommodation is created horizontally in the same shifting direction of the new equilibrium profile. However, they ignore the role of tectonics in foreland basin as the cause of vertical (flexural) movements and equilibrium profile disruption that occurs along the upstream portion of the fluvial
system. Flexure mostly induces creation of vertical accommodation not horizontal, as indicted by our models.

Non-marine foreland basins, especially along the upstream portions of the fluvial systems (the case in the Putumayo Basin), are controlled by the graded profile, whose disruption creates or destroy accommodation. Nomenclature regarding periods of equilibrium profile disrupted by tectonic uplift (basin deepening), or periods dominated by sediment-related deflection (long-distance cratonward onlap) would result more appropriate. Wedge-like geometry of sedimentary packages not only represents period of tectonic activity, which only produces ~ 20% of total subsidence in Putumayo, but also reflects the distribution of vertical loads.

Observations in Putumayo indicate that tectonic pulses may be represented by onlap shifts in the hinterland direction that indicate basin deepening (Figs. 4.19, 4.20 and 4.25). The predicted lense-like geometry during tectonic quiescence is not present in the Putumayo area. Tectonic quiescence probably is represented by onlap advance in the cratonic direction. Hiatuses might represent periods of fluvial-system equilibrium or graded stage. This is one mechanism, that using flexure, explains the development of fluvial sedimentary cycles and the creation of regional hiatuses in areas non-affected by faulting or folding, like in the Putumayo (Summary of model in Figure 4.25).

Sandstone/shale ratios grow with increasing deflection (subsidence) rate during Rumiyaco, Orteguaza-Belen-Orito and Ospino depositional time (Fig. 4.26), in contrast with the theory of Allen (1978), and might indicate that the sedimentation of coarse-grained sediments are related to an increasing rather than to a decreasing rate of
subsidence. An increase in sandstone/shale ratio with subsidence rate is consistent with creation of relief between the depocenter and sediment source.

Tectonic-related subsidence produces an initial geometry of the basin that is later enlarged (up to 35%) and distorted by sedimentary loads. Therefore, the position and geometries of forebulges are difficult to identified using stratigraphic pinch outs. Elevation of these protuberances is subtle and difficult to distinguish from relief produced by authigenic processes.

The effective elastic thickness of the lithosphere in the Putumayo Basin found in this study agrees with the ranges of values found in previous studies and in nearby areas (Stewart and Watts, 1997; Ojeda, 2001; Sarmiento, 2001; Watts, 2001). Similar to previous studies in the Cretaceous retro-arc foreland along Idaho-Wyoming area (Dickinson, 1974) and the Alps peripheral foreland (Homewood, et al., 1986; Sinclair et al., 1991) the flexure of the lithosphere can be modeled using elastic models with constant strength. Lateral changes in the effective elastic thickness are not necessary for modeling the basin, within the studied area.

Composed forebulges appear to move back and fore randomly during the evolution of the basin. Tectonic-related forebulges however move in a more systematic way, since they move forward towards the foreland (craton) or remain stationary, but do not move backwards. This indicates a mostly forward breaking thrust sequence history.

The effective elastic thickness does not change at a scale of $10^7$ m.y., according to the results obtained in this study, similar to the results of Sinclair et al. (1991) in the Alps. No viscous-elastic relaxation seems to be necessary to explain the flexure in the Putumayo Basin. Neither the curvature nor the sedimentary cover, via blanketing (Lavier
and Steckler, 1997) are sufficient to weaken the plate during the evolution of the basin. Cretaceous back-arc rifting process did not affect the lithosphere in Putumayo, therefore the thermal age of the plate is probably Precambrian to Cambrian (Sarmiento, 2001). The effective elastic thickness found in this work is similar to, or slightly lower than the estimated thickness of the crust in the area (35-40 km) (Sarmiento, 2001).

During the latest episode of the Andean orogeny in Colombia in late Miocene to Recent times, the subsidence rate reaches 144 m/m.y. This might indicate that this collisional event was stronger than the previous ones. However, it is during this last tectonic event when the models show the lowest values of flexure caused by tectonic loading in the basin (between 17 and 10) and indicate an increase in the volume sediment flux related to this orogeny.

**Summary**

There is no evident relationship between tectonosedimentary units and eustatic third order cycles in Putumayo basin. Subsidence caused by tectonic loading is one order of magnitude higher than the expected eustatic fluctuations, whereas basement subsidence cause by sediment loading is up two orders of magnitude higher.

Seismic expression of foreland deposits consist of mostly onlap geometries, with some basinward shifting events that reflect basin deepening. Downlap reflector geometry, toward the basin depocenter is also distinguishable and may represent periods of sediment sourcing from the foreland (craton). It is advisable not to use the passive-margin sequence nomenclature in non-marine foreland basins, since there are not trangressive-regressive events.
Figure 4.25 Conceptual model (not at scale) of sedimentary sequences in foreland basins. Each tectonostratigraphic cycle in Putumayo is thought to be developed in a similar process. Accommodation is initially induced by the thrusting that deflects the plate (Figs. 4.23 and 4.24). Later sediments enlarge and deepen the deflection of the sedimentary basement until the river system reaches its equilibrium stage and a hiatus develops. A new tectonic pulse is marked by onlap shift towards the thrust belt.
Effective elastic thickness of the continental lithosphere underlying the Putumayo Basin in Colombia is 20 to 40 km. It did not change during the ~ 65 m.y. basin history. About 400 km of continental lithosphere were flexed during the evolution of the basin. The maximum curvature of the plate was attained during Miocene time. It was not sufficient to weaken the plate.

Sediment loading is responsible for up to 78% of flexure in one particular cycle (Miocene to Holocene). The maximum amount of tectonic loading was created during Oligocene time, it reaches 26% of the total deflection. Deflection rate in Putumayo varies between 63 and 140 m/m.y. The average is ~ 85 m.y.
The geometry that the Putumayo basin exhibits today is largely due to the wide sediment loads (up to 400 km). The sediment-induced deflection obscures the initial deflection produce by tectonic loading. As a result, forebulges are not distinguishable from seismostratigraphic records, probably due to their subtle topography.

Fluvial sedimentary systems can develop stratigraphic sequences during the adjustment of the system profile to its graded stage. Disruption of the equilibrium profile induced upstream by tectonic loading and plate-flexure creates accommodation. Then, sediments enlarge, deepen and fill the basin until the graded stage is regain. During this period hiatuses develop before a new flexural-induce creation of accommodation reinitiate.

References


Daly, M. C., 1989. Correlations between Nazca/Farallon Plate kinematics and forearc basin evolution in Ecuador. Tectonics, v. 8, pp. 769-790.


Mora, C., Torres, M. P. and Escobar, J., 1997. Potencial generador de hidrocarburos de la Formación Chipaque y su relación estratigráfica secuencial en la zona axial de la


CHAPTER 5. CONCLUSIONS

Effective elastic thickness does not change in foreland basins during their evolution in both Timor Sea peripheral basin and Putumayo retroarc basin. This implies that the strength of the lithosphere is probably inherited and last over periods of 5 m.y. (Timor Sea) to 65 m.y. (Putumayo). It also implies that the lithosphere behaves elastically. Contrary to Lavier and Steckler (1997) sediment cover did not thermally isolate the plate and have not effect on the strength of the underlying lithosphere. Subsidence related to tectonic loading reaches up to 70% in Timor Sea, a sediment-starving basin. Present day bathymetry of the Timor Trench, a very narrow and depth depression, is produced in part by lack of sedimentation. By contrast in Putumayo, the tectonic-related subsidence reaches up to 25%, however the basin is enlarged and deepened by sedimentary loads that account for the remaining 75% of subsidence. Eustasy is responsible for only 2 to 3% of the creation of accommodation space in Timor (a marine basin). In Putumayo, the fluvial system that operated during the evolution of the basin is not affected by changes of the base level induced by eustatic fluctuations. It rather responds to changes in equilibrium profile that in turn are controlled by flexure. The scale of eustatic changes is, at least one order of magnitude less than subsidence rates created by supracrustal loads. Therefore changes in space for sedimentation in foreland basins is due mostly to flexure or movements of the base of the accommodation space.

Mechanically, both Timor Sea and Putumayo basins produce qualitatively similar results and deflections evolve at rates of similar order of magnitude. Effective elastic thickness estimates show that Putumayo (20 to 40 km) is weaker that Timor Sea (70 to 100) even thought the lithosphere in Putumayo is older (~ 500 Ma) than the Australian
lithosphere (~250 Ma). This can be explained invoking the models of Burov and Diament (1995). They predict that in a decoupled lithosphere, is the crust that supports the vertical stresses, therefore the plate will appear weak, which could be the explanation for the strength values obtained in Putumayo. In Timor Sea, by contrast, the Australian lithosphere appears to be strong. It probably indicates a coupled plate, i.e. there is mechanical continuity between the crust and the mantle portions of the lithosphere. However, no other modeling was carried out in this study to address this difference, therefore this explanation requires additional support. The difference in plate strength is also responsible for the difference in basin width between Timor (~500 km) and Putumayo (~ 400 km). Forebulges, as seem today, are composed structures developed after overlapping of bulges created by tectonic and sedimentary loads at different time and space. They do not represent limits for sedimentation. Regionally, a sedimentary pinch-out does not represent the position of the forebulges; pinch outs might have different causes in foreland basins.

References


APPENDIX 1. FLEXURAL THEORY

This appendix is a summary of the flexural theory taken from two books: *Geodynamics: Applications of Continuum Physics to Geological Problems* (Malloy Eds.) by D. Turcotte and G. Schubert (1982), and *Isostasy and Flexure of the Lithosphere* (Cambridge University Press) by T. Watts (2001). Some sentences are paraphrased from these two books. Although this is not an original approach, it is necessary to understand the implemented models used in this study.

A plate of thickness \( h \) and width \( L \) pinned at its ends that bends under the load of a line force \( V_a \) (Nm\(^{-1}\)) applied at its center, is the mechanical framework used for flexural models. The plate is infinitely long in the \( z \) direction, and is assumed to be thin compared with its width, \( h \ll L \), and the deflection of the plate \( w \) is taken to be small, \( w \ll L \). This assumption is necessary to justify the use of linear elastic theory (Fig. A1.1).

The deflection of a plate can be determined by requiring it to be in equilibrium under the action of all forces and torques exerted on it.

![Figure A1.A. Model of deflection of an elastic beam pinned at both ends (After Turcotte and Schubert, 1982).](image)
The forces and torques on a small section of the plate between horizontal locations \( x \) and \( x + dx \) (Fig. A1.2)

![Diagram showing forces and torques on a small section of a plate]

Figure A1.2. Distribution of forces and torques on a small portion of the plate (After Turcotte and Schubert, 1982).

A downward force per unit area \( q(x) \) is exerted on the plate by whatever distributed load the plate is required to support, and an upward force exerted by the foundation that acts at every point in proportion to the deflection (Winkler Foundation) per unit area equal to \( t(x) \).

The downward load per unit length in the \( z \) direction, between \( x \) and \( x + dx \) is \( q(x)dx \). The shear force \( V \), per unit length in the \( z \) direction, acts on the cross section of the plate normal to the plane of the figure (it is the resultant of all shear stresses integrated over a cross-sectional area of the plate). The horizontal force \( P \), per unit length in the \( z \) direction, is assumed to be independent of \( x \).

The net bending moment \( M \), per unit length in the \( z \) direction, is the integrated effect of the moments exerted by the normal stresses \( \sigma_{xx} \) (also called fiber stress).
At location \( x \) along the plate the shear force is \( V \), the bending moment \( M \) and the deflection \( w \). At \( x + dx \), the shear force is \( V + dV \), the bending moment is \( M + dM \) and the deflection is \( w + dw \). All \( V, M \) and \( P \) are per unit length in the \( z \) direction. A force balance between \( x \) and \( x + dx \) produce

\[ q(x)dx + V - (V + dV) - t(x) = 0 \]

Then,

\[ \frac{dv}{dx} - t(x) = -q(x) \]

or \[ \frac{d(dM)}{dx(dx)} - t(x) = -q(x) \Rightarrow \frac{d^2M}{dx^2} + q(x) = t(x) \]

The moment balance between \( x \) and \( x + dx \) requires that \( V \) and \( V + dV \) produce a net torque \( Vdx \) (clockwise sense). Horizontal forces \( P \) are small compared with vertical forces, therefore they are can be neglected. The balance moment is given by:

\[ M - (M + dM) + (Vdx) = 0 \]

\[ \frac{dM}{dx} = V \]

or

\[ \frac{d(dM)}{dx(dx)} - t(x) = -q(x) \]

or

\[ \frac{d^2M}{dx^2} + q(x) = t(x) \quad (1) \]

If the plate is deflected downwards (Fig. A1.3) the upper half of the plate is contracted and the longitudinal stress \( \sigma_{xx} \) is positive; the lower half is extended and the
longitudinal stress $\sigma_{xx}$ is negative. The fiber stress $\sigma_{xx}$ is zero on the mid-plane $y = 0$, which is the neutral unstrained surface.

![Figure A1.3. Fiber stress model for a bended elastic plate (After Turcotte and Schubert, 1982).](image)

The net effect of these stresses on the plate is a counterclockwise bending moment on the cross section of the plate.

The force on an element of the plate’s cross-section of thickness $dy$ is $\sigma_{xx}dy$. This force exerts a torque about the midpoint of the plate given by $\sigma_{xx}ydy$. If the torque over the cross section of the plate is integrated, the bending moment ($M$) is obtained.

$$M = \int_{-h/2}^{h/2} \sigma_{xx}ydy$$

Stress is accompanied by a longitudinal strain $\varepsilon_{xx}$ that is positive in the upper half and negative in the lower part. Strain and stress can be related using Young’s modulus (E) and Poisson’s ratio ($\nu$):

$$\varepsilon_{xx} = \frac{1}{E} (\sigma_{xx} - \nu \sigma_{zz})$$
\[ \varepsilon_{zz} = \frac{1}{E} (\sigma_{zz} - v \sigma_{xx}) \]

Then,

\[ \sigma_{xx} = \frac{E}{(1-v^2)} \varepsilon_{xx} \]

\( M \) becomes,

\[ M = \frac{E}{(1-v^2)} \int_{-h/2}^{h/2} \varepsilon_{xx} y \, dy \]

The longitudinal strain \( \varepsilon_{xx} \) depends on the distance from the mid-plane \( y \) and the local radius of curvature \( R \) (Fig. A1.4).

In the figure A1.5 \( \phi \) is \( l/R \) in radians. The change in length of the section \( \Delta l \) at a distance \( y \) from the mid-plane is:

\[ \Delta l = -y \phi = -y \frac{1}{R} \]

\[ \varepsilon_{xx} = \frac{\Delta \lambda}{l} = \frac{y}{R} \]

Figure A1.4 Radius of curvature concept (After Turcotte and Schubert, 1982).
The radius of curvature $R$ is determined by the change in slope of the plate mid-plane with horizontal distance (Fig.A1.5).

If $w$ is small $-dw/dx$, the slope of the mid-plane is also the angular deflection of the plate from the horizontal $\alpha$. The small angle $\phi$, is the change in $\alpha$, that is, $d\alpha$ is the small distance $l$ or $dx$. Thus,

$$\phi = d\alpha = \frac{d\alpha}{dx} dx = \frac{d}{dx}(-\frac{dw}{dx})dx = \frac{d^2w}{dx^2} dx$$

Then:

$$\frac{1}{R} = \frac{\phi}{l} = \frac{\phi}{dx} \approx -\frac{d^2w}{dx^2}$$

The strain is given by:

$$\varepsilon_{xx} = -y\frac{d^2w}{dx^2}$$

And then the bending moment is
\[ M = \frac{-E}{(1-v^2)} \int_{-h/2}^{h/2} y \, dy \]

Then,

\[ M = \frac{-E h^3}{12(1-v^2)} \frac{d^2 w}{dx^2} \]

The coefficient of this equation is called the flexural rigidity \( D \).

\[ D \equiv \frac{E h^3}{12(1-v^2)} \]

The bending moment is the flexural rigidity divided by the curvature:

\[ M = -D \frac{d^2 w}{dx^2} = \frac{D}{R} \]

Substituting this equation in equation (1)

\[ D \frac{d^4 w}{dx^4} - q(x) = -t(x) \quad (2) \]

This is the general equation for flexural deflection of elastic beams under vertical loads in an inviscid fluid.

The net force acting on the lithosphere \( q \) can be calculated by assuming that there is an equilibrium between the surface load \( q(x) \) and the upward hydrostatic force per unit area \( t(x) \). The restoring hydrostatic force per unit area \( t(x) \) is equivalent to the force that results from replacing mantle rock of thickness \( w \) and density \( \rho_m \) by the infill material (water or sediments) of thickness \( w_f \) and density \( \rho_f \):

\[ t(x) = (\rho_m - \rho_f) g w \]

Replacing \( t(x) \) in equation 2, produces

\[ D \frac{d^4 w}{dx^4} + (\rho_m - \rho_f) g w = q(x) \]
Which is the general equation used for the deflection of the lithosphere under supracrustal loads. The model of the deflection of the oceanic or continental plates assumes a linear load $V_o$ at $x=0$ (Fig. A1.6).

![Deflection model for an oceanic plate with a linear load](image)

Figure A1.6. Deflection model for an oceanic plate with a linear load (After Turcotte and Schubert, 1982).

Since the applied load is zero except at $x=0$, so the general equation becomes:

$$D \frac{d^4w}{dx^4} + (\rho - \rho_0)gw = 0$$

The general solution for this equation is

$$w = e^{x/\alpha} \left( A \cos \frac{x}{\alpha} + B \sin \frac{x}{\alpha} \right) + e^{-x/\alpha} \left( C \cos \frac{x}{\alpha} + D \sin \frac{x}{\alpha} \right)$$

Where $A$, $B$, $C$ and $D$ arbitrary integration constants. $\alpha$ is a parameter that determines the amplitude and wavelength of the deflection and is given by:

$$\alpha = \left[ \frac{4D}{(\rho - \rho_0)} \right]^{1/4}$$

The integration constants, considering beams of infinite length under a line load (figure above), are found using the following boundary conditions:

1. $x \to \pm\infty, z \to 0$ This condition ensures that there is no deflection at the far edges of the beam. The condition is met if $e^{x/\alpha}$ vanishes in equation (3). This occurs if $A=B=0$ in equation (3) and reduces to
where load is applied. These conditions are met if the first derivative of the deflection curve (Eq. 3.1; Fig, A1.7) is zero at $x=0$:

$$\frac{dy}{dx} = C(-\frac{1}{\alpha} e^{-x/\alpha} \cos \frac{x}{\alpha} - \frac{1}{\alpha} e^{-x/\alpha} \sin \frac{x}{\alpha}) + D(-\frac{1}{\alpha} e^{-x/\alpha} \sin \frac{x}{\alpha} + \frac{1}{\alpha} e^{-x/\alpha} \cos \frac{x}{\alpha})$$

At $x=0$, the equation is only zero if $C=D$. Equation 3.1 becomes

$$w = e^{-x/\alpha} D(\cos \frac{x}{\alpha} + \sin \frac{x}{\alpha}) \quad (3.2)$$

The remaining integration constant $D$ can be determined by considering the force balance; i.e. when the downward acting load equal the sum of all upward acting loads:

$$q(x) + 2\rho_{\gamma} g \int_{0}^{x} ydx = 2\rho_{m} g \int_{0}^{x} ydx$$

The factor 2 ensures that both side of the load are counted. This equation becomes:
\[ q(x) = 2(\rho_m - \rho_{\text{eff}}) g \int_0^\infty y \, dx \]

Then

\[ q(x) = 2\alpha (\rho_m - \rho_{\text{eff}}) gD \]

Substituting this equation in equation 3.2 gives:

\[
\begin{align*}
w &= \frac{q(x)}{2\alpha (\rho_m - \rho_{\text{eff}}) g} e^{-x/\alpha} \left( \cos \left( \frac{x}{\alpha} \right) + \sin \left( \frac{x}{\alpha} \right) \right) \\
&= \frac{2q(x)}{\alpha (\rho_m - \rho_{\text{eff}}) g} e^{-x/\alpha} \left( \cos \left( \frac{x}{\alpha} \right) \right)
\end{align*}
\] (4)

For the case of a semi-infinite beam with a free end, the deflection is still given by equation 3, with the constants of integration yet to be determined. Since the plate extends from \( x = 0 \) to \( x = \infty \) and it is required that \( w = 0 \) as \( x \to \infty \), \( A \) and \( B \) must be again zero. It is assumed that no external torque is applied to the end \( x=0 \). From \( \frac{d^2w}{dx^2} = 0 \) at \( x=0 \), this requires that \( D= 0 \). Equation 3.1 becomes

\[ w = e^{-x/\alpha} (C \cos \frac{x}{\alpha}) \] (3.2)

Since the vertical shear stress at the end \( x=0 \) is equivalent to the applied vertical load then

\[ q(x) = \frac{\alpha (\rho_m - \rho_{\text{eff}}) gC}{2} \]

Equation 3.2 becomes

\[
\begin{align*}
w &= \frac{2q(x)}{\alpha (\rho_m - \rho_{\text{eff}}) g} e^{-x/\alpha} \left( \cos \left( \frac{x}{\alpha} \right) \right)
\end{align*}
\]

Following Hetenyi (1946), a formula can be derived for situations in which the linear load (for a semi-infinite beam) is loaded within the end \( x=0 \) (Fig. 1A8).

Equation 4 becomes:
This represents the deflection of an elastic plate under a linear load at any position $x$ along the plate. This equation is used in the MatLab routine (Appendix 4).

![Diagram of deflection for an elastic beam loaded within.](image)

Distributed loads can be divided in individual linear loads $q(x)$, whose individual deflection can be integrated to obtain the resultant deflection $\Sigma w(x)$. This method is called superposition by Hetenyi (1946). This allows to determined deflection for a variety of geometry of the load, as shown in the following figure A1.9:
Figure A1.9. Model of distributed load and superposition (After Hetenyi, 1946).

References
APPENDIX 2. VERTICAL LOADING BY TECTONIC WEDGE AND SEDIMENTS

Tectonic Wedge

Critical taper theory (Davis et al., 1983) states that tectonic wedges (thrust belts) behave as a Coulomb material (brittle), i.e. the wedge is a granular aggregate whose fragments frictionally slide past one another. The wedge geometry results from a relative displacement produced by horizontal stresses and from gravitational sliding. The gravitational energy, due to elevation of the hinterland, contributes to both vertical and horizontal stresses as well. As the backstop moves, the wedge deforms internally (via faults, folds and penetrative strain, Fig. A2.1), as a consequence its slope increases. When the wedge reaches a critical taper ($\phi_c$) (define as the surface slope ($\alpha$) plus the detachment dip ($\beta$)), the wedge slide stably toward the foreland along the basal detachment.

During foreland translation of the wedge, new material is added to the wedge at its toe by break-forward thrusting, causing the taper angle to decrease. When this happens, stable sliding stops, and internal deformation occurs once again within the wedge by back or out-of-sequence thrusting. The observed critical taper reflects this balance and depends on the material strength of the wedge. In thrust-belt mechanics it is assumed that the taper angle is maintained throughout deformation history of the wedge.

In this work, the tectonic loading is represented by a thrust-belt (tectonic wedge), and the total vertical stress that causes deflection is equivalent to the amount of loading that comes from forward modeling. Therefore, it is possible distribute these values along the wedge as a function of the critical taper angle.
Figure A2.1. Mechanical model of the Coulomb-wedge theory (After van der Pluijm and Marshak, 1997).
The 2D representation of the tectonic wedge is a triangular area (Fig. A2.2), whose dimensions can be calculated by:

\[ h = x \tan \phi_c \]

Where \( \phi_c \) is the critical taper angle, \( h \) is the elevation of the wedge in meters at the interior-thicker end of the load and \( x \) is the length in meters of the wedge (Fig. A2.2).

The vertical force \( P_o \) is the summation of discrete linear loads of length \( dx \),

\[ dP_o = \rho g \tan \phi_c x dx \]

Where \( \rho \) is the density of the wedge material, and \( g \) is the gravitational acceleration. Integration of equation 2 along \( x \) derives in:

\[ P_o = \rho g \tan \phi_c \frac{x^2}{2} \]

Then

\[ x = \sqrt{\frac{2P_o}{\rho g (\tan \phi_c)^2}} \]

For the MatLab implementation, the value of \( x \) is divided in discrete 1 km long rectangles of different height (\( h \)), for easy computer implementation. Each rectangle is assumed to behave as an independent linear load and the superposition method is used.

**Sediment Loading**

Following the same discretization criteria used in the tectonic-wedge approach, the sediment record, taken from seismostratigraphic data, is divided into 10 km-wide rectangles, each of which acts a an individual linear load. The approach to the wedge geometry of the sediments is used because the MatLab routine calculates the thickness of the column as a function of the slope angle. This angle (\( \theta \)) is measures from the
decompacted sediment thickness. The lateral extent of the sedimentary cover comes from the seismic data and geological constraints.

Figure A2.2. Model of the tectonic distributed load implemented in the computer code. \( x \) is horizontal distance, \( h \) is height and \( q \) the amount of loading

The decompacted thickness is the vertical load applied to the lithosphere after deflection caused by tectonic loading. The superposition method is used to determine the total deflection of the plate.

Figure A2.3. Model of the sedimentary distributed-load implemented in the computer code
References

APPENDIX 3. DATA REDUCTION-DECOMPACTION

It is assumed that compaction is the mechanical and irreversible process of porous-water expulsion from void space in sedimentary rocks. In order to determine the thickness of any stratigraphic unit at depositional time (decompacted thickness) it is necessary to know the present day porosity.

Well data provides the most reliable source of information for present day porosities. In absent of well information it is possible to use the equation derived by Athy (1930) and Hedberg (1936) for normal pressure:

\[ \phi = \phi_0 e^{-cz} \]

Where \( \phi \) is present day porosity, \( \phi_0 \) is the surface porosity, \( Z \) is depth and \( c \) is the slope of the curve; it changes according to the lithology (Sclater and Christie (1980) and the basin.

Decompacted thickness is then calculated using Watts (1993):

\[ S_i^* = \frac{s_i (1 - \phi_{si})}{(1 - \phi^*_{Si})} \]

Where \( S_i^* \) is the present day thickness, \( S_i \) is the decompacted thickness of the layer, \( \phi_{si} \) is the present day porosity \( \phi^*_{Si} \) is the porosity of the decompacted layer.

The density of the decompacted sediment is given by:

\[ \rho_{si}^* = \rho(\phi^*) + \rho_{gr} [1 - (\phi^*_{si})] \]

Where \( \rho_{si}^* \) is the density of the decompacted layer, and \( \rho_{gr} \) is the sediment grain density.
Rate of sedimentation $R$ is the vertical accumulation of sediments by unit time and is defined by:

$$R = \frac{S_i}{T}$$

Where $T$ is elapsed time or time duration.

References


APPENDIX 4. MATLAB ROUTINE

%---------------------------------------------------------------------------------------------------------
%Constants and equation for broken plate
%---------------------------------------------------------------------------------------------------------

%=[Input Data
%---------------------------------------------------------------------------------------------------------
g= 10 ;%Gravity acceleration m/s2 []
Psw= 2300 ;%density of tectonic wedge material kg/m3 []
Pw= 1035 ; %water density kg/m3 []
Ps= 1800; %decompacted sediment density kg/m3 []
Pm= 3300 ;  %density mantle kg/m3 []
E= 5e11 ; % Young Modulus in Pa (kg/m-s2) []
v= 0.25 ; % Poisson ration []

%---------------------------------------------------------------------------------------------------------

% EET km, alpha and lambda calculations
%---------------------------------------------------------------------------------------------------------

Te=100000; %effective elastic thickness in meters []
D= [(E*Te^3)/(12*(1-v^2))]; %Calculation of flexural rigidity (N-m)80 km/EET
Alpha=[(4*D)/((Pm-Pw)*g)]^2.5 ; % Determination of flexural parameter
Lambda=1/Alpha ; % Characteristic length--> Hetenyi

%---------------------------------------------------------------------------------------------------------
% Horizontal array

\[ X = [0:10000:130000]'; \] horizontal distance in meters, with intervals of 10000 meters; (position of the load with respect to \( x=0 \) )

% ESTIMATION OF THE LOAD ALONG THE SEDIMENT WEDGE N/m

\[ a = (000000:10000:600000)'; \] position left of load

\[ z = (\text{max}(a) - \text{min}(a):-10000:0)' \] \% ARRAY OF THE WEDGE

\[ \Psi = 0.2 \] \% Slope angle. Measured from sediment decompacted thickness

\[ h = \text{tan}(\Psi/180) \times z; \] Elevation of the load-column

\[ dp = Ps \times g \times h \times 10000; \] \% linear load along the wedge

\[ P = \text{sum}(dp); \] \% summation of loading

% CONVOLUTION OF SEDIMENT WEDGE

\[ \text{for } j = 1: \text{size}(X) \text{ \% for values } X \]

\[ F(j) = \text{Lambda} \times X(j); \] \% array of values using lambda and \( x \) to easier calculations

end

\[ \text{for } k = 1: \text{size}(a) \text{ \% for values of } a \]

\[ Fa(k) = \text{Lambda} \times a(k); \] \% array for load at 'a'

end
%---------------------------------------------------------------------------------------------------------
% FUNCTION SOLUTIONS FOR SEDIMENT WEDGE
%---------------------------------------------------------------------------------------------------------
cosF= cos(F) ;%cos of F
sinF=sin(F) ;%sin F
expF=exp(F) ;%e^F
cosFa= cos(Fa) ;%cos of F for a
sinFa=sin(Fa) ;%sin F for a
expFa=exp(Fa) ;%e^F for a

%---------------------------------------------------------------------------------------------------------
% QUADRATIC SOLUTIONS FOR FLEXURAL EQUATION FROM HETENYI (1946)
%---------------------------------------------------------------------------------------------------------
for r=1:size(X)
    Dz(r)=(exp(-F(r))).*(cosF(r)) ; % Functions for origin conditions
    Az(r)=exp(-F(r)).*[cos(F(r))+sin(F(r))] ;
    Bz(r)=exp(-F(r)).*sin(F(r)) ;
    Cz(r)=exp(-F(r)).*[cos(F(r))-sin(F(r))] ;
end

for s=1:size(a)
Dza(s)=(exp(-Fa(s))).*(cosFa(s)) ; % Functions for origin conditions for 'a'

Aza(s)=exp(-Fa(s)).*[cos(Fa(s))+sin(Fa(s))] ;

Bza(s)=exp(-Fa(s)).*sin(Fa(s)) ;

Cza(s)=exp(-Fa(s)).*[cos(Fa(s))-sin(F(s))] ;

end

Wta = zeros(size(X),size(a));

for w=1:size(X) ; %for z
    for y=1:size(a) ; % for a
        Wta(w,y)= [(Cza(y)+(2*Dza(y))).*Az(w)-[2*(Cza(y)+Dza(y)).*Bz(w)]+Az(abs(y-w)+1)]; %eq. 23a Hetenyi (1946)
    end
end

%-------------------------------------------------------------------------------------------------------------------------------------

% FLEXURAL EQUATION FOR SEDIMENT LOADING
%-------------------------------------------------------------------------------------------------------------------------------------

Flexure = Wta*dp/((Pm-Pw)*g*2*Alpha); % Final flexure due to sediments (matrix multiplication)

%-------------------------------------------------------------------------------------------------------------------------------------

% PLOTTING SEDIMENTARY LOADING
%-------------------------------------------------------------------------------------------------------------------------------------
xa=[min(a):10000:(max(X)+min(a))]; %Graphical array for sediment load

hh=[size(xa)]; %Graphical array for geometry of sediment load

hh=zeros(hh);

hh(1:size(h))=h; %Graphical representation of sediment load; hold on

% EUSTASY ADJUSTMENT []

Delta_Eustasy=0; % []

if Delta_Eustasy < 0 ;
    Acc=Flexure + Delta_Eustasy;
elseif Delta_Eustasy >= 0;
    Acc=Flexure+(Delta_Eustasy/.71);%Eustatic compensation. Assume local isostasy
end

% ESTIMATION OF TECTONIC LOAD ASSUMING IT BEHAVES AS A COULOMB MATERIAL

% at=(0:1000:10000)'; %position left of load []

zt =(max(at):-1000:0)' ;%ARRAY OF THE WEDGE []

Theta= 6; %Critical taper angle for tectonic wedge []

ht=tan(Theta*pi/180)*zt; %high of tectonic load
dpt=g*Psw*ht*1000; % linear load along the wedge

Pt=sum(dpt); % summation of the load along the wedge n/m

%---------------------------------------------------------------------------------------------------------
% CONVOLUTION OF TECTONIC WEDGE
%---------------------------------------------------------------------------------------------------------

for j=1:size(X) % for values X
  Ft(j)= Lambda*X(j); % array of values using lambda and x to easier calculations
end

for k=1:size(at) % for values of a
  Fat(k)=Lambda*at(k); % array for load at 'a'
end

%---------------------------------------------------------------------------------------------------------
% FUNCTION SOLUTIONS FOR TECTONIC WEDGE
%---------------------------------------------------------------------------------------------------------

cosFt= cos(Ft); % cos of F
sinFt=exp(Ft); % e^F

cosFat= cos(Fat); % cos of F for a
sinFat=sin(Fat); % sin F for a
expFat=exp(Fat); % e^F for a
for r=1:size(X)
Dzt(r)=exp(-Ft(r)).*cosFt(r) ; % Functions for origin conditions
Azt(r)=exp(-Ft(r)).*[cos(Ft(r))+sin(Ft(r))];
Bzt(r)=exp(-Ft(r)).*sin(Ft(r));
Czt(r)=exp(-Ft(r)).*[cos(Ft(r))-sin(Ft(r))];
end

for s=1:size(at)
Dzat(s)=exp(-Fat(s)).*cosFat(s) ; % Functions for origin conditions for 'a'
Azat(s)=exp(-Fat(s)).*[cos(Fat(s))+sin(Fat(s))];
Bzat(s)=exp(-Fat(s)).*sin(Fat(s));
Czat(s)=exp(-Fat(s)).*[cos(Fat(s))-sin(Fat(s))];
end

Wtat = zeros(size(X),size(at));%flexure
for w=1:size(X) ; %for z
    for y=1:size(at) ; % for a
        Wtat(w,y)= [(Czat(y)+(2*Dzat(y))).*Azt(w) -
        [2*(Czat(y)+Dzat(y)).*Bzt(w)]+Azt(abs(y-w)+1)]; %eq. 23a Hetenyi (1946)
    end
end
end;

%-----------------------------------------------------------------------------------------------------------------------
% FLEXURAL EQUATION FOR TECTONIC LOADING
%-----------------------------------------------------------------------------------------------------------------------
Flexure_Tec = Wtat*dpt/((Pm-Pw)*g*2*Alpha); % Final flexure matrix multiplication

%-----------------------------------------------------------------------------------------------------------------------
%DEFINING TECTONIOC LOADING ARRAY
%-----------------------------------------------------------------------------------------------------------------------
hht=[size(X)];
hht=zeros(hht);
hht(1)=max(ht);
hht(3)=min(ht);

%-----------------------------------------------------------------------------------------------------------------------
%SEDIMENT EXCESS CALCULATION
%-----------------------------------------------------------------------------------------------------------------------
SEX= hh - Acc'; %Amount of sediment excess. accommodation space due to Sediment and Eustasy
SEX(SEX<=0)=0 ;% For Thickness < than Acosta.
ToDe=(Acc + SEX'); %flexure represented by sediment loading and sediment excess (Tectonic)=thickness (To be model)
Total_Flexure=(Acc + Flexure_Tec); %Total flexure. Summation of tectonic +
Sediments + Eustasy

% PLOTTING

plot (X,Flexure_Tec,'k'); hold on
set(gca, 'ydir','reverse')
xlabel('Horizontal Distance (meters)')
ylabel('Deflection (meters)')
title ('Tectonic Deflection- Seismic data-window')

%END

References
Well penetration chart of Putumayo basin. Data gathered by Ecopetrol and Geotec and reported by Cordoba et al. (1997). Unconformities are found in areas deformed by thrusting. Areas without deformation representation hiatuses. According to Cordoba et al. (1997) this chart is based on data from well-cores and extensive paleontological studies.
References
VITA

John obtains his bachelor’s degree in geology from the Universidad Nacional de Colombia, in Bogotá, in 1995, after, and in spite of, five years among soccer fanaticism, liberal bohemians and romantic socialists. Soon after his graduation, while working for oil and raw-material companies and as well as a consultant, he begins to map part of the Colombian rain forest, the Amazon, and the Caribbean coast. One and a half years of long-walks, multiple and unbearable helicopter flights, nights in the dark jungle, a lot of rain and experiences with bizarre plants, snakes, tarantulas, monkeys, psychedelic birds and beautiful purple dragonflies, led him to move to a more civilized environment. Using London as a base, he spent one year trying to learn some english and back-packing Europe. Getting back again, to a more rural life, he earns his Master’s degree in applied geophysics from the University of Alabama in 2000. The following year, he enrolls at LSU as doctoral student in geosciences. After this scholastic period, following the advise of his father, he has decided to devote himself (at least for a while) to a more sedentary and contemplative life, dividing his time between exploring and exploiting oil, and reading contemporary literature and about the history of geology.